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New England Intercollegiate Geological Conference  
81st Annual Meeting

Guidebook for Field Trips in

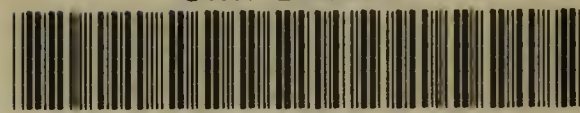
**SOUTHERN  
AND  
WEST-CENTRAL  
MAINE**

October 13, 14 and 15, 1989

Hosted by the  
Department of Sciences and Mathematics  
University of Maine at Farmington  
Farmington, Maine

Editor  
Archie W. Berry, Jr.





# New England Intercollegiate Geological Conference

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## TABLE OF CONTENTS

FIELD TRIPS	Page
A-1    Geology and Geochemistry of the Mt. Agamenticus Complex, York, Maine <i>John Brooks, David Gust, and Arthur Hussey</i> .....	1
A-2    Illinoian and Late Wisconsinan Till in Eastern New England: a Transect from Northeastern Massachusetts to West-Central Maine <i>T.K. Weddle, B.D. Stone, W.B. Thompson, M.J. Retelle, D.W. Caldwell,</i> <i>and J.M. Clinch</i> .....	25
A-3    Hydrogeology and Development Pressures, Brunswick, Maine <i>Peter Garrett, Donald Newburg</i> .....	86
A-4    The Chain Lakes Massif and its Tectonic History <i>Eugene L. Boudette, Gary M. Boone, Richard Goldsmith</i> .....	98
A-5    Significance of Al-silicate in Staurolite-Grade Rocks, Central Maine <i>Michael Holdaway, Barbara Dutrow</i> .....	122
B-1    Mineralogic and Textural Evidence for Polymetamorphism along a Traverse from Phillips to Oquossoc, Maine <i>C. V. Guidotti</i> .....	144
B-2    Glacial Geology of the Androscoggin River Valley in Oxford County, Western Maine <i>Woodrow B. Thompson</i> .....	165
B-3    Melanges and Turbidite Facies of the Madrid Formation, Central Maine <i>Dwight Bradley, Lindley Hanson</i> .....	183
B-4    Continuation of A-4 .....	
B-5    Environmental Geology Along the Sandy River, Farmington, Maine <i>Tom Eastler, Joel Sproul, Andy Buckland</i> .....	200
C-1    Repeat of A-1 .....	
C-2    Repeat of B-2 .....	
C-3    Repeat of A-3 .....	
C-4    Bedrock Geology of the Southeastern Margin of the Connecticut Valley-Gaspe Synclinorium in Somerset County, Maine <i>Robert G. Marvinney</i> .....	212
C-5    Repeat of B-5 .....	

THEORY OF THE EARTH

1. The Earth as a System	1
2. The Earth's Surface	2
3. The Earth's Interior	3
4. The Earth's Atmosphere	4
5. The Earth's Hydrosphere	5
6. The Earth's Biosphere	6
7. The Earth's Geosphere	7
8. The Earth's Lithosphere	8
9. The Earth's Pedosphere	9
10. The Earth's Atmosphere	10
11. The Earth's Hydrosphere	11
12. The Earth's Biosphere	12
13. The Earth's Geosphere	13
14. The Earth's Lithosphere	14
15. The Earth's Pedosphere	15
16. The Earth's Atmosphere	16
17. The Earth's Hydrosphere	17
18. The Earth's Biosphere	18
19. The Earth's Geosphere	19
20. The Earth's Lithosphere	20
21. The Earth's Pedosphere	21
22. The Earth's Atmosphere	22
23. The Earth's Hydrosphere	23
24. The Earth's Biosphere	24
25. The Earth's Geosphere	25
26. The Earth's Lithosphere	26
27. The Earth's Pedosphere	27
28. The Earth's Atmosphere	28
29. The Earth's Hydrosphere	29
30. The Earth's Biosphere	30
31. The Earth's Geosphere	31
32. The Earth's Lithosphere	32
33. The Earth's Pedosphere	33
34. The Earth's Atmosphere	34
35. The Earth's Hydrosphere	35
36. The Earth's Biosphere	36
37. The Earth's Geosphere	37
38. The Earth's Lithosphere	38
39. The Earth's Pedosphere	39
40. The Earth's Atmosphere	40
41. The Earth's Hydrosphere	41
42. The Earth's Biosphere	42
43. The Earth's Geosphere	43
44. The Earth's Lithosphere	44
45. The Earth's Pedosphere	45
46. The Earth's Atmosphere	46
47. The Earth's Hydrosphere	47
48. The Earth's Biosphere	48
49. The Earth's Geosphere	49
50. The Earth's Lithosphere	50
51. The Earth's Pedosphere	51
52. The Earth's Atmosphere	52
53. The Earth's Hydrosphere	53
54. The Earth's Biosphere	54
55. The Earth's Geosphere	55
56. The Earth's Lithosphere	56
57. The Earth's Pedosphere	57
58. The Earth's Atmosphere	58
59. The Earth's Hydrosphere	59
60. The Earth's Biosphere	60
61. The Earth's Geosphere	61
62. The Earth's Lithosphere	62
63. The Earth's Pedosphere	63
64. The Earth's Atmosphere	64
65. The Earth's Hydrosphere	65
66. The Earth's Biosphere	66
67. The Earth's Geosphere	67
68. The Earth's Lithosphere	68
69. The Earth's Pedosphere	69
70. The Earth's Atmosphere	70
71. The Earth's Hydrosphere	71
72. The Earth's Biosphere	72
73. The Earth's Geosphere	73
74. The Earth's Lithosphere	74
75. The Earth's Pedosphere	75
76. The Earth's Atmosphere	76
77. The Earth's Hydrosphere	77
78. The Earth's Biosphere	78
79. The Earth's Geosphere	79
80. The Earth's Lithosphere	80
81. The Earth's Pedosphere	81
82. The Earth's Atmosphere	82
83. The Earth's Hydrosphere	83
84. The Earth's Biosphere	84
85. The Earth's Geosphere	85
86. The Earth's Lithosphere	86
87. The Earth's Pedosphere	87
88. The Earth's Atmosphere	88
89. The Earth's Hydrosphere	89
90. The Earth's Biosphere	90
91. The Earth's Geosphere	91
92. The Earth's Lithosphere	92
93. The Earth's Pedosphere	93
94. The Earth's Atmosphere	94
95. The Earth's Hydrosphere	95
96. The Earth's Biosphere	96
97. The Earth's Geosphere	97
98. The Earth's Lithosphere	98
99. The Earth's Pedosphere	99
100. The Earth's Atmosphere	100



The Geology and Geochemistry of the Agamenticus Complex, York, Maine

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## INTRODUCTION

The Triassic Agamenticus Complex is the earliest phase of subalkalic to alkalic magmatism associated with the opening of the Atlantic Ocean during the Mesozoic Era. Recent geochemical analyses and field mapping have helped to clarify the petrologic relationships between syenite and granite phases present within this complex. Several features of this plutonic body make it an interesting focus of study: 1) the complex is the earliest pulse of rift-related magmatism in New England 2) the complex contains syenites and granites with no clear evidence of a basaltic parent, 3) gravity and magnetic anomalies indicate that such a parent is not present in the near surface, and 4) portions of the pluton have undergone significant deuteritic metasomatism.

## GENERAL

Topographic sheets: 15'- York and Kennebec; 7.5'- York Harbor, York beach, Wells, and North Berwick.

The Agamenticus Complex is located in southwestern Maine in the coastal townships of York, Wells and South Berwick. This is a region of low to moderate relief often covered with dense vegetation. Portions of the complex form topographic highs and provide good bedrock exposures. Elsewhere within the complex outcrops are best exposed along a limited number of secondary roads, small stream beds, and glacially oversteepened valley walls. A nearly continuous N-S cross section of the complex is provided by the outcrop exposures along I-95, but, permission must be obtained from the Maine State Turnpike Authority to visit these outcrops.

## REGIONAL SETTINGS

### Paleozoic Lithologies

The Agamenticus Complex intrudes the Precambrian to Ordovician Kittery and Eliot Formations of the Merrimack Group and the Devonian Webhannet Pluton (Hussey, 1962, 1985; Osberg et al., 1985; Gaudette et al., 1982) (Fig. 1). The Kittery Formation is comprised of thin- to thick-bedded feldspathic and calcareous quartzites, quartzites, siliceous phyllites, with subordinate interlayered marble beds. The Eliot Formation contains thin interbedded phyllite and quartzous phyllite. Rock types within the Webhannet Pluton range from quartz diorite to biotite granite.

The rocks of the Kittery and Eliot Formations experienced regional greenschist facies metamorphism prior to the emplacement of the Ordovician Exeter Diorite (Hussey, 1985; Gaudette et al., 1982). Contact metamorphism of these formations is apparent in close proximity to the Agamenticus Complex and in foundered blocks within the complex.

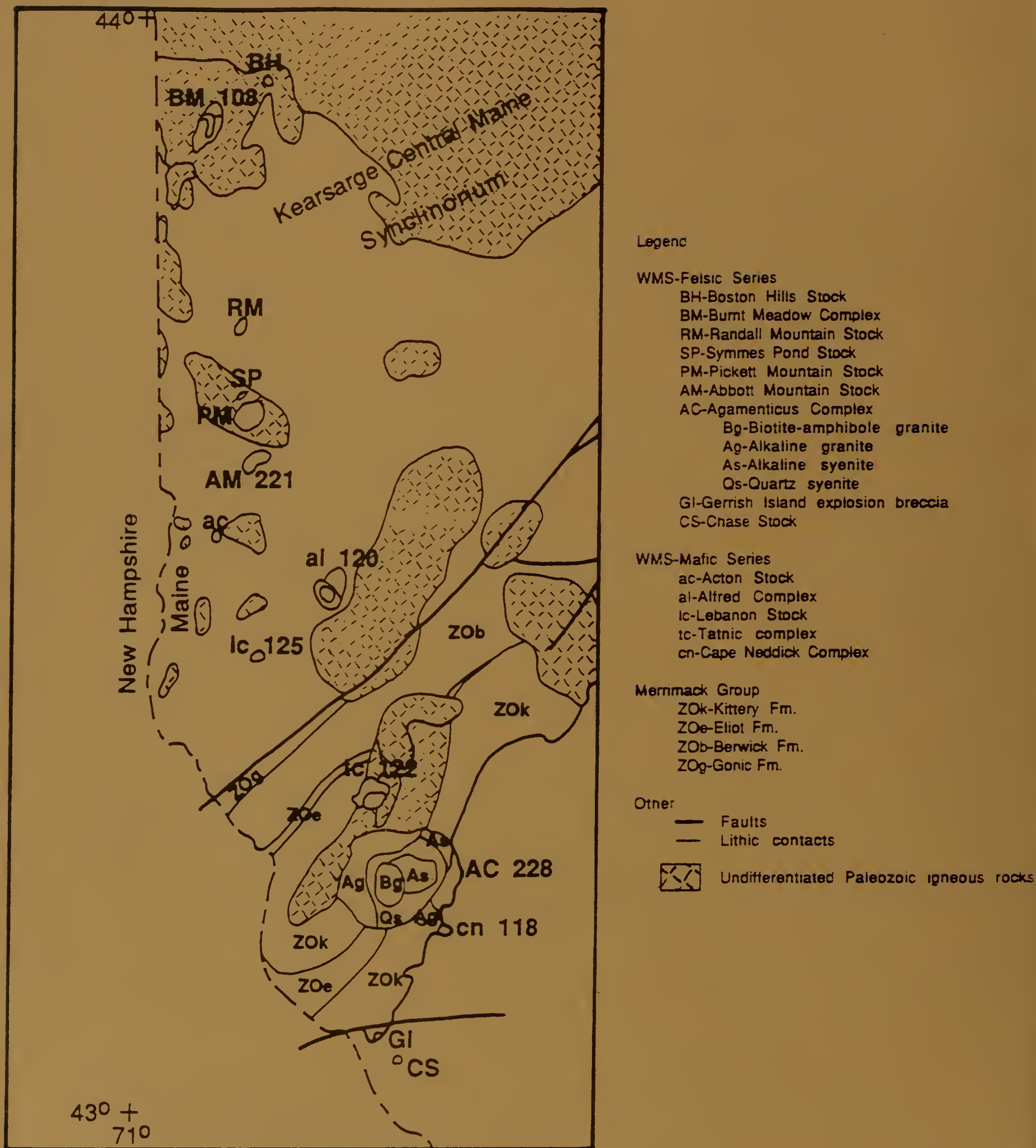


Fig. 1- Simplified geology of southwestern Maine. Adapted from Hussey 1985. Ages (Ma) indicated where known (Foland and Faul, 1977; Foland et al., 1977; Hoefs, 1967). Except for the Merrimack Group the metamorphic rocks are undifferentiated.



Xenoliths and foundered blocks of calcareous quartzite of the Kittery Formation contain abundant epidote and diopside indicating upper amphibolite facies metamorphism.

## Mesozoic Plutons

The opening of the Atlantic Ocean was preceded and accompanied by the intrusion of the Mesozoic, subalkalic to alkalic White Mountain Plutonic Series (WMS) (Billings, 1956; Bedard, 1965; McHone and Butler, 1985). The Agamenticus Complex is at the southern end of a NNW-trending belt of small (1 to 50 km<sup>2</sup>), felsic or mafic WMS bodies in southwestern Maine (Gilman, 1972, 1979; Hussey, 1962, 1985; Osberg et al., 1985). Two of these bodies, the mafic Tatnic and Cape Neddick Complexes, are located in the vicinity of the Agamenticus Complex (Fig. 1). The gabbros to granodiorites of the Tatnic and Cape Neddick Complexes were intruded during the early Cretaceous period (122 to 119 Ma) (Foland and Faul, 1977) and are the last pulse of rift-related magmatism in this area. The felsic Agamenticus Complex, intruded during the late Triassic period (216-228 Ma) (Foland and Faul, 1977; Foland et al., 1971; Hoefs, 1967), is the earliest stage of rift-related magmatism.

## AGAMENTICUS COMPLEX

### Overview

Mapped originally by Wandke (1922), the Agamenticus Complex was later remapped in reconnaissance studies by Woodard (1957) and Hussey (1962, 1985). On the basis of this work, the complex was subdivided into four major lithologies (Fig. 1); alkaline syenite, alkaline granite, porphyritic biotite-amphibole granite, and "contaminated alkaline granite" (later renamed quartz syenite) (Hussey, 1962, 1985). From the oldest to the youngest, the relative ages of the phases established by cross-cutting relationships and textural arguments are alkaline syenite, alkaline granite, and porphyritic biotite-amphibole granite. The quartz syenite is interpreted by Hussey (1962) as the product of variable degrees of assimilation of the syenitic phase by the intruding alkaline granite.

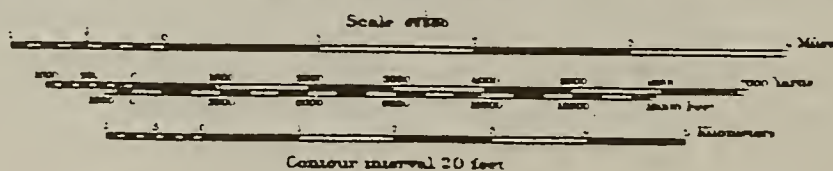
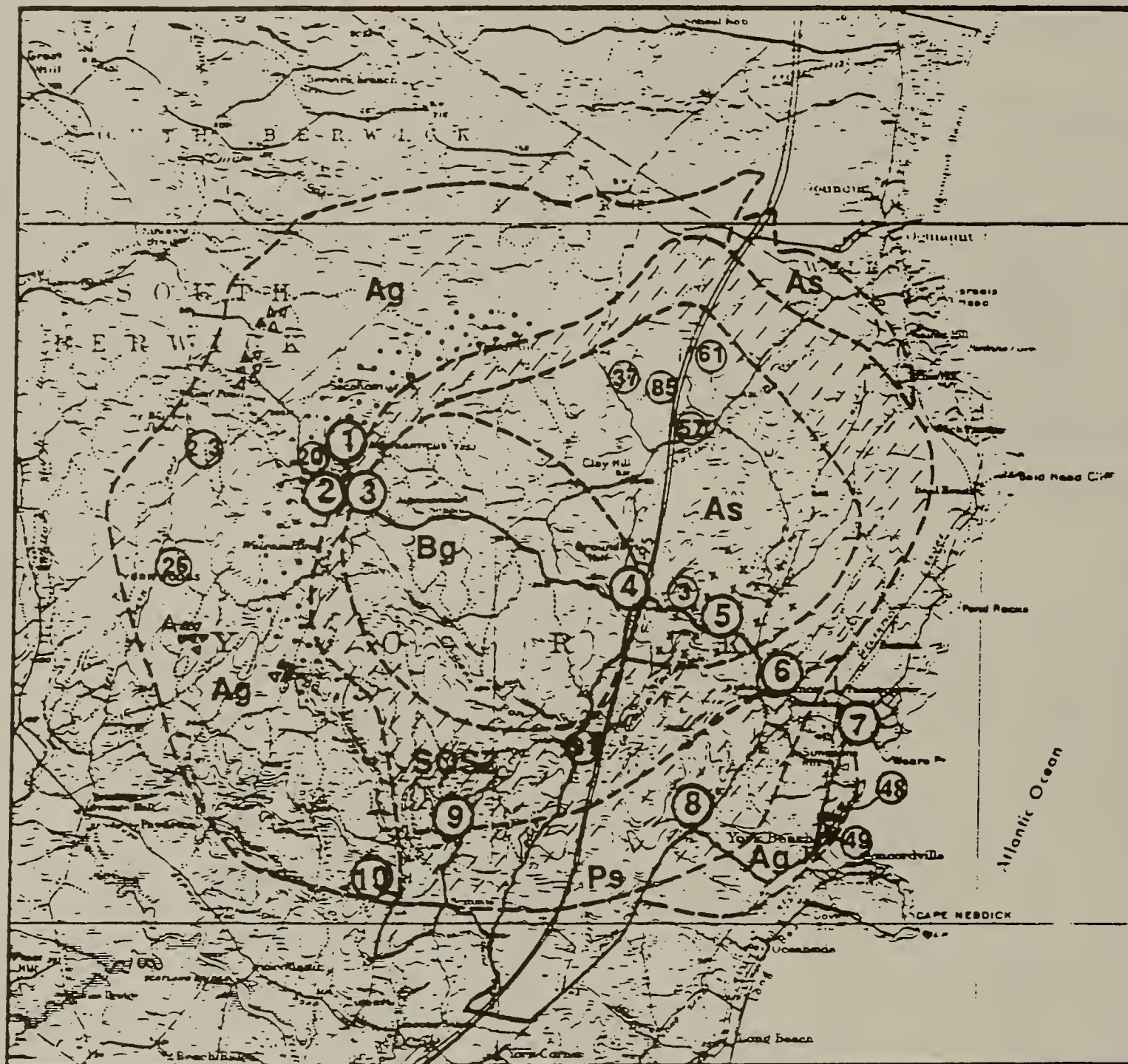
As part of ongoing research on the White Mountain Plutonic Series in southwestern Maine, the Agamenticus Complex was recently mapped at a scale of 1:24000. Major, trace and REE analyses on selected samples were conducted using XRF, ICP, and INAA methods. Mineral chemistry from a limited number of samples was obtained by electron microprobe analysis. The findings of current mapping substantiate the general phase relations as previously mapped by Hussey (1962). However, the more detailed mapping conducted in this research has shown that broad regions of textural and mineralogical variability are observed within individual phases. The "contaminant zone" has been subdivided into an aenigmatite-bearing syenite unit and a syenite to quartz syenite zone. The geology shown in Figure 2 reflects these changes in detail and in interpretation. We have changed the modifier 'alkaline' to alkalic to reflect current petrologic diction.

### Alkalic granite to quartz syenite

Although the contacts of the western lobe of the alkalic granite remain unchanged, field work revealed a significant amount of leucocratic quartz syenite within the eastern portion of the body (Fig. 2; Table 1). Contacts between the alkalic granite and quartz syenite have not been observed and the two phases are considered to be transitional. Xenolith-rich areas occur within the alkalic granite (Fig. 2) and indicate that these regions were possibly close to the roof of the magma chamber. In xenolith zone A, amphiboles are subophitic to ophitic rather than interstitial as is typical of the alkalic granite elsewhere. This could reflect an increase in fluid pressure towards the top of the magma chamber during



Fig. 2 - Revised geology of the Agamenticus Complex. The starting point of the field trip is located just south of map coverage.



#### Legend

##### Agamenticus Complex

Bg- Biotite-amphibole granite

Ag- Alkalic granite

-quartz syenite portion of the alkalic granite

As- Amphibole-bearing alkalic syenite

-alkalic syenite

Ps- Porphyritic aenigmatite syenite

SQSZ- Syenite to quartz syenite zone

-Zone of pervasive intrusion by alkalic granites

-Xenolith rich zone

##### Other

-- Lithic contacts

— Field trip route

③ Stop numbers

④⑨ Sample location numbers other than stops

Table 1 - Mineral modes for representative rock types within the Agamenticus Complex.

Sample#	MAG5	MAG20	MAG4	MAG4	MAG4	MAG11b	MAG139	MAG3	MAG 91-1
Stop #	2		4	4	4	9	8		
Rock type	alk gran	alk gran	bio-amb	gran	mafic xeno	porph sye	porph sye	sye	monzo-sye
Mineralogy									
ksp	70.58	61.72	27.96			70.10	69.51	85.88	70.46
plg	1.40	4.17	31.65		44.22	5.24	5.30	4.86	8.68
qtz	12.07	31.86	31.65		2.26	6.95	6.16	1.38	1.52
amb	15.37	2.00	2.99			16.29	6.34		5.74
aenlg						1.24	5.49		
blo		0.08	5.52		50.25				
cpx									
olv									
other*	0.58	0.17	0.23		3.27	0.19	0.57	4.60	11.72
sum	100.00	100.00	100.00		100.00	100.00	100.00	100.00	100.00
*	Zr,Ap,Op,Al	Op,Fl	Zr,Ap,Op,Al			Op	Ap,+ Op	Op,Ap	Op,Ap

Abbrv.: ksp-alkali feldspar,plg-plagioclase,qtz-quartz,amb-amphibole,aenig-aenigmatite,bio-biotite,  
cpx-clinopyroxene,olv-olivine,Zr-zircon,Ap-apatite,Op-opaque,Fl-fluorite,Al-allanite



crystallization, the greater impact of deuteritic alteration, or the presence of a separate pulse of alkalic granite. The arfvedsonite amphibole compositions (Fig. 3) observed at this location support the interpretation of enhanced deuteritic alteration (Giret et al., 1980) but the subophitic to ophitic textures suggest that amphibole growth was synchronous with the growth of potassium feldspar.

The eastern lobe of alkalic granite (Fig. 1 and 2) is well exposed along the coastline adjacent to York Beach. At this locality the alkalic granite is medium- to coarse-grained and commonly contains miarolitic cavities, granitic autoliths, and xenoliths of Kittery Formation. An approximately 6 meter wide, northeast-trending dike or segregation of very coarse-grained to pegmatitic granite occurs within the central portion of the outcrop. Amphibole-rich trachytes intrude as 6 cm to 5 m wide dikes and contain abundant xenoliths of alkalic granite and Kittery Formation. These trachytic dikes are considered to be coeval with the late stage trachytic dikes that cut across all of the other phases within the complex.

#### Porphyritic biotite-amphibole granite

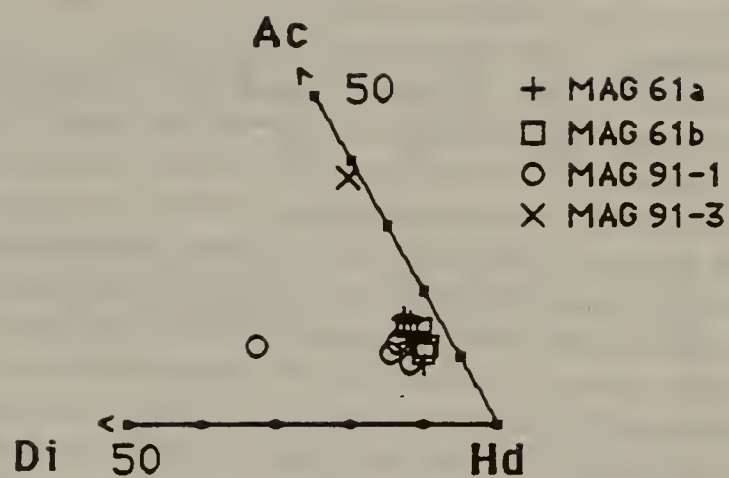
This rock is a gray to pink, fine- to medium-grained, porphyritic, subsolvus granite (Fig. 1 and 2; Table 1). The phenocryst assemblage is comprised of plagioclase and orthoclase and the matrix contains, quartz, biotite and subordinate amphibole. Textural variations of both matrix grain size and phenocryst abundance are observed within and between outcrops. These textural variations and the presence of Kittery-type xenoliths suggest that the current level of exposure was near the side or roof of the original magma chamber.

#### Alkalic Syenite

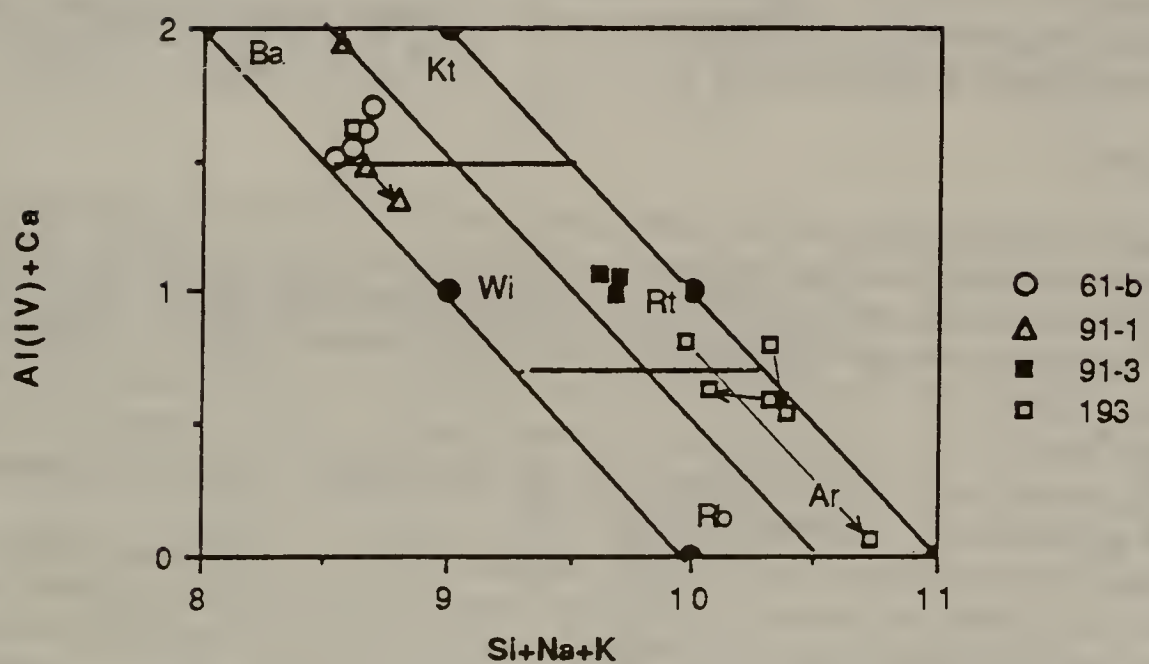
The alkalic syenite varies considerably in texture and in mineralogy (Fig. 2). The northern portions of the alkalic syenite are coarse to medium-grained and the southern portion is fine-grained. Perthite is the major phase in the alkalic syenite with interstitial hedenbergite, fayalite, and quartz occurring in the southern portions (Mag 3, Table 1). Other areas of the alkalic syenite contain the same mineralogy plus barrositic amphibole. The amphibole is interstitial and often rims the pyroxene. Amphibole compositions indicate either late magmatic crystallization, minor deuteritic alteration (Giret et al., 1980), or both in this localized area. However, a contact observed between the medium-grained, amphibole-bearing syenite (Mag 61b, Fig. 2) and a fine-grained syenite mineralogically similar to the southern alkalic syenite suggests that the amphibole-bearing syenite instead may be a separate, more hydrous magma pulse.

#### "Contaminant Zone"

Field relationships within the "contaminant zone" are complex and are not yet fully understood. This complexity has resulted from the multiple intrusion in this zone of several syenite pulses with the subsequent intrusion by one or more granitic magmas. High volatile contents resulted in the growth of sodic and iron (Fe<sup>3+</sup>)-rich amphibole, turbid perthitic feldspars, interstitial microcline, the reaction of primary phenocrysts, and the development of greisen zones. Intrusion style within this zone was brittle to plastic-brittle as seen by the sharp angularity of many of the xenoliths. Some xenoliths have been rounded by mechanical disaggregation during transport and many have undergone variable degrees of assimilation. Ghost-like xenoliths of melanocratic to mesocratic syenites to quartz syenites are surrounded by alkalic granite. We suggest that the efficiency of assimilation in this zone is due to 1) the similarity in composition between the various syenites and the alkalic granite and 2) the likelihood that the syenites, although solid, were quite warm and that the difference in temperature between the syenites and the alkalic granite was relatively small.



a)



b)

Fig. 3 - a) Clinopyroxene compositions plotted on an Acmite, Diopside, Hedenbergite (Ac-Di-Hd) ternary diagram (after Mitchell and Platt, 1977). b) Amphibole compositions for the Agamenticus Complex. Arrows connect core to rim compositions. Ba- Barrosite, Rt- Richterite, Wi- Winchite, Rb- Riebekite, Ar- Arfvedsonite (after Giret et al., 1980).



Within the southern and southeastern portion of the complex a porphyritic aenigmatite (cossyrite)-bearing syenite is observed (Fig. 2; Table 1). This syenite may be one of the earliest intrusions in the complex as it is intruded by both the eastern (Stops 6 and 8), and western lobes (Stop 9), of the alkalic granite. It occurs as a major phase in some outcrops and as disaggregated blocks in the alkalic granite in others. Another syenite (Mag 91-1, Table 1), located on I-95 between the alkalic syenite and the aenigmatite-bearing syenite is distinctive as it contains coarse-grained plagioclase and significant amounts of pyroxene. As can be seen in Table 1, modal proportions of plagioclase and alkali feldspar of this syenite are different from the main body of alkalic syenite. Both feldspars occur as inclusion-free phenocrysts with rims of potassium feldspar. Plagioclase phenocrysts are strongly zoned. Pyroxene is more magnesium-rich (Fig. 3) and plagioclase is more calcic than comparable phases in the alkalic syenite and suggest that this early pyroxene syenite is more primitive than the alkalic syenite. As discussed previously, amphibole as well as the potassium feldspar rims on plagioclase probably reflect deuteric alteration.

The complexity of rock types within the "contaminant zone" or quartz syenite (Hussey, 1962, 1985) and the process that was responsible for its creation make it the least understood but petrologically most exciting part of the Agamenticus Complex. To portray the geological variety of this zone with respect to a regional map, we have renamed it the syenite to quartz syenite (SQSZ) unit. This unit does not include the porphyritic syenite described above. In the geologic map that accompanies the text (Fig. 2) a slash pattern is added to this zone, and to the aenigmatite-bearing syenite, to emphasize that this region is a contact zone which contains a significant proportion of intrusive alkalic granite. The width of the contact suggests that it is subhorizontal in the southern portion of the complex and more steeply dipping in the north.

## Geochemistry

Major element abundances, CIPW norms, and DI's ( $1/3 \text{ Si} - (\text{Ca} + \text{Mg} + \text{Mn})$ ) for representative samples from the Agamenticus Complex are given in Table 2. All phases, with the exception of the mafic syenites, exhibit low abundances of CaO, MgO, MnO and TiO<sub>2</sub> and moderate amounts of K<sub>2</sub>O and Na<sub>2</sub>O. Mafic syenites, present at a limited number of localities (MAG 57D, 125CP (Stop 10)), have slightly lower SiO<sub>2</sub> and higher Al<sub>2</sub>O<sub>3</sub>, CaO, MgO, MnO, TiO<sub>2</sub>, and P<sub>2</sub>O<sub>5</sub> than the other syenites. A major inflection occurs on many of the major, and to a lesser extent trace element, Harker-type variation diagrams (Fig. 4) at approximately 65 weight percent SiO<sub>2</sub>.

Rocks of the Agamenticus Complex are dominantly silica-saturated to silica-oversaturated. Acmite and sodium silicate occur as normative peralkalic minerals within a number of the SQSZ phases, the aenigmatite-bearing syenite in the southern portion of the complex, and the western lobe of the alkalic granite. The only silica-undersaturated rocks within the complex are the mafic syenites, containing 3 to 4 weight percent normative nepheline. The porphyritic biotite-amphibole granite is corundum normative (peraluminous). Plotted on a Qtz-Ab-Or ternary diagram (Fig. 5), the syenites and alkalic granites lie along the 1 to 3 kbar thermal trough for low An contents and moderate PH<sub>2</sub>O. The porphyritic biotite-amphibole granites are slightly more orthoclase-rich.

The rocks of the Agamenticus Complex are generally more agpaitic than other WMS rocks (Fig. 6). Many of the other WMS bodies, some of which contain mafic members, evolve from miaskitic through pulmaskitic to agpaitic compositions (Fig. 6) and from metaluminous to peraluminous and peralkalic compositions. On the agpaitic plot (Fig. 6) the mafic syenites plot within the pulmaskitic field and the other rocks plot within the agpaitic field and exhibit a rough correlation of increasing agpaitic index to decreasing  $(\text{K} + \text{Na})/(\text{Si}/6)$ . The porphyritic biotite-amphibole granite is distinct from this trend and



Table 2 - Major element analyses (Wt percent), CIPW normative assemblages, and Differentiation Indexes (DI) for selected samples.

SAMPLE #	Sye	Amb sye	B+h granite		Alk Gran-West lobe		Alk gran-East lobe		
	MAG-2	MAG 39	MAG-4	MAG-179	MAG-5	MAG-23	MAG-26	MAG-48	MAG-49
STOP #	5		4	3	2				
SiO2	65.7	61.7	72.24	71.5	69.34	75.64	70.95	73.8	70.9
TiO2	0.46	0.63	0.41	0.39	0.37	0.09	0.24	0.18	0.33
Al2O3	15.88	17.5	14	14.14	15.17	13.2	14.43	9.79	11.8
FeO	0	0	0	0	0	0	0	0	0
Fe2O3	6.52	5	3.23	2.62	5.59	0.63	3.12	5.6	5.08
MnO	0.2	0.12	0.07	0.05	0.15	0.03	0.07	0.08	0.09
MgO	0.27	0.5	0.7	0.5	0.25	0.13	0.12	0.1	0.1
CaO	1.56	1.87	1.4	1.08	0.91	0.21	0.53	0.26	0.34
Na2O	5.68	5.45	2.95	3.87	5.42	4.77	5.6	4.67	4.93
K2O	5.82	6.26	6.36	5.2	5.5	4.47	5.27	4.17	4.39
P2O5	0.05	0.15	0.13	0.15	0.07	0.02	0.02	0.05	0.05
LOI		0.6		0.56		0.33	0.23	0.43	1.16
SUM	102.14	99.18	101.49	99.5	102.77	99.19	100.35	98.7	98.01
CIPW Normative assemblages (FeO/Total Fe= .85)									
Or	35.26	38.18	31.41	31.48	33.78	26.74	31.83	26.37	27.76
Ab	49.29	47.61	31.99	33.56	46.61	40.86	45.88	29.05	38.8
An	0.33	4.95	6.12	4.49	3.14	0.92			
Q	3.74		24.93	25.42	9.78	30.18	16.43	34.41	25.08
C			0.59	1.11		0.37			
Di	5.48	2.89			0.96		2.01	0.81	1.14
Hy	3.11	1.48	2.7	2.27	3.42	0.57	1.73	3.87	3.33
Ol		2.17							
Nb									
Ac									
Ns								2.3	2.36
AP	0.11	0.34	0.34	0.34	0.18	0.04	0.04	2.4	0.74
Il	0.9	1.24	0.93	0.76	0.91	0.17	0.47	0.12	0.12
Mt	1.46	1.13	0.69	0.58	1.22	0.14		0.37	0.67
DI	25.14	24.67	28.47	27.8	26.73	30.09	28.53	28.65	28.09

Table 2 contd - Major element analyses (Wt percent), CIPW normative assemblages, and Differentiation Indexes (DI) for selected samples.

SAMPLE #	Melano- to mesocratic SQSZ MAG-11a	MAG-91-1	Felsic SQSZ MAG-11c	Mafic syenites MAG-57D	MAG-125CP	Aenig sye MAG-139
	9		9		10	8
SiO2	63.9	60.18	67.9	57.93	59.22	62.11
TiO2	0.45	0.82	0.09	0.84	0.86	0.54
Al2O3	14.6	14.78	16.7	17.22	17.65	14.58
FeO	0	0	0	0	0	0
Fe2O3	7.69	9.59	2.12	6.65	6.73	8.76
MnO	0.2	0.24	0.02	0.14	0.14	0.22
MgO	0.1	0.43	0.1	1.95	1.97	0.13
CaO	1.54	2.89	0.13	3.27	3.28	1.46
Na2O	6.22	5.87	6.48	6.29	5.94	6.34
K2O	4.94	4.64	5.71	4.53	4.56	5.24
P2O5	0.06	0.22	0.05	0.23	0.25	0.09
LOI	0.2	0.15	0.39	0.27	0.46	0.3
SUM	99.7	99.66	99.3	99.05	100.6	99.47
CIPW Normative assemblages (FeO/Total Fe= .85)						
Or	30.42	28.35	35.15	27.14	27.22	33.45
Ab	53.29	51.37	56.14	44.97	44.06	49.52
An		0.29		7.4	6.72	
Q	0.21		4.48			
C						
Di	5.75	10.32	0.36	6.06	6.63	5.64
Hy	4.22	0.78	1.37			4.68
Ol		4.62		8.6	7.71	0.39
Ne				3.53	4.06	
Ac	3.64		0.88			4.11
Ns	1.38		1.34			0.89
AP	0.16	0.5	0.07	0.52	0.51	0.21
Il	0.93	1.61	0.2	0.32	1.64	1.11
M1		2.16		1.46	1.45	
DI	24.01	20.03	29.05	17.24	17.33	23.9

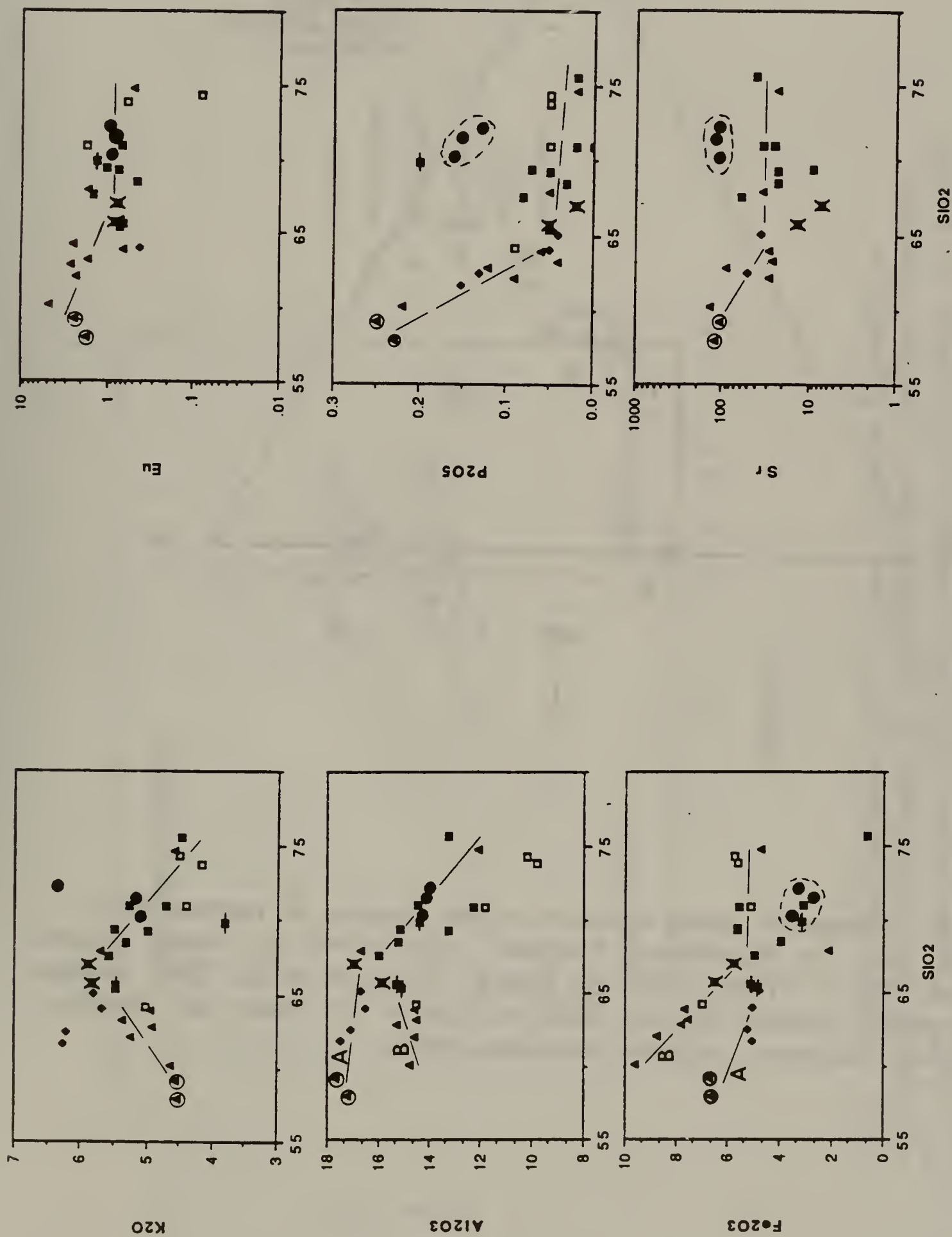


Fig. 4 - Harker diagrams of selected major and trace elements. A flexure commonly occurs at approximately 65% SiO<sub>2</sub> on these diagrams. On the Al<sub>2</sub>O<sub>3</sub> and Fe<sub>2</sub>O<sub>3</sub> the rocks of the SQSZ plot along a separate trend (B) from the mafic and amphibole-bearing syenites (A). Biotite-amphibole granites often plot separately from the other phases. Symbols: X - alkalic syenite, filled circle - amphibole-bearing alkalic syenites, filled square - mafic syenites, filled triangle - biotite-amphibole granites, filled diamond - misc. dikes, filled circle - SQSZ.



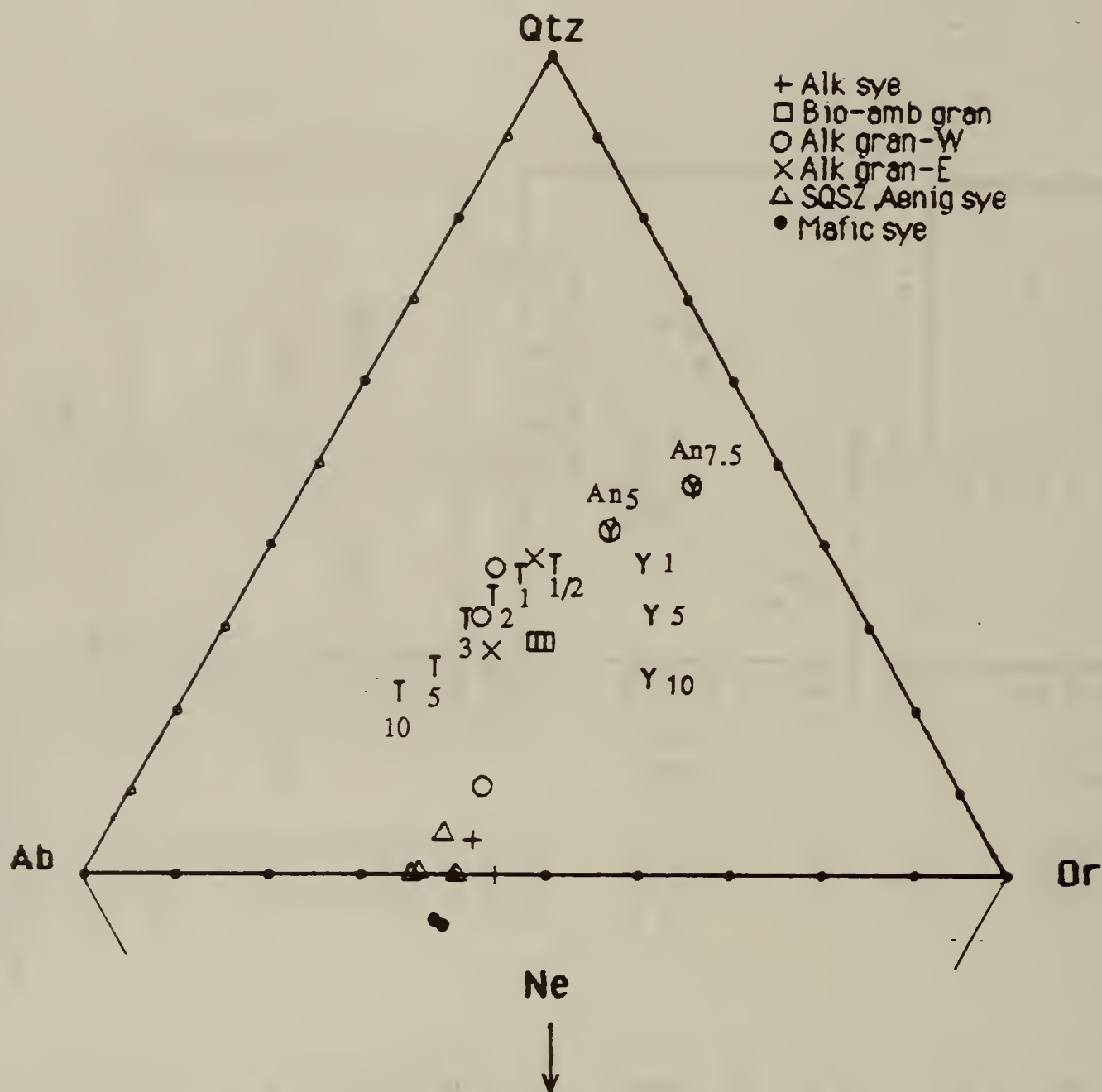


Fig. 5 - Normative quartz-albite-orthoclase diagram of representative samples from the Agamenticus Complex. Also shown are ternary minima (  $\tau$  ) for  $PH_2O = P_{Total}$  (Tuttle and Bowen, 1958; Luth et al., 1964), minima in An-bearing systems (  $\oplus$  ) and  $PH_2O = P_{Total} = 1kb$  (James and Hamilton, 1969), and anhydrous minima (  $\gamma$  ) (Luth, 1969).

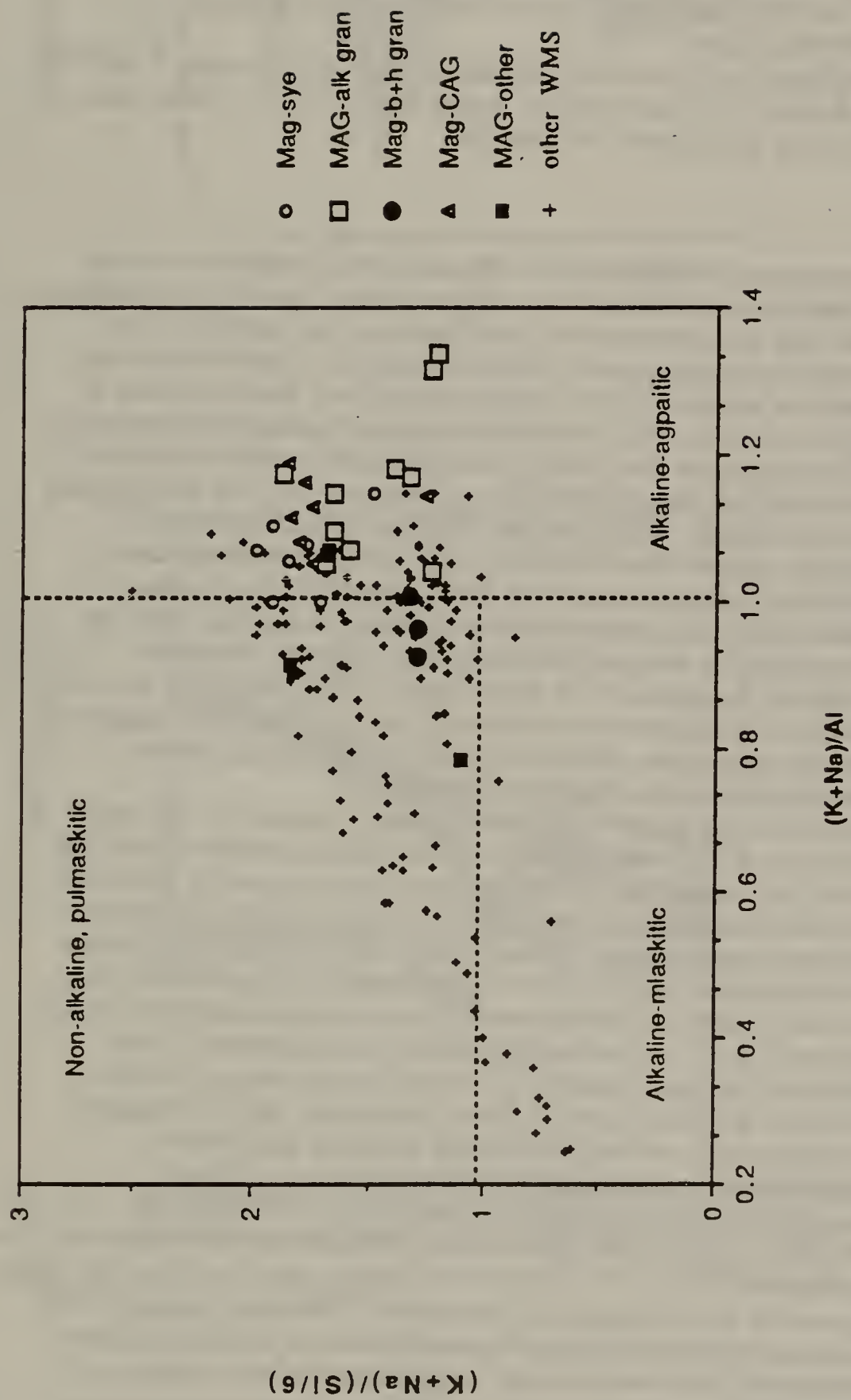


Fig. 6 - Agpaitic compositional plot of rocks from the Agamenticus complex and other WMS bodies in southwestern Maine and New Hampshire. Plot shows the evolutionary trend from miaskitic mafic phases to agpaitic felsic phases. The porphyritic biotite-amphibole granite plots within a discrete field from the main evolutionary trend.



plots within a well-defined field of pulmaskitic/agpaitic WMS rocks. This discrete field lies slightly below the main evolutionary trend exhibited by mafic to felsic WMS rocks.

Trace and rare earth element abundances for representative samples are presented in Figure 7. With the exception of Sr and Ba, the different phases all have similar trace element patterns. The large ionic lithophile (LIL) (except Sr) and high field strength (HFS) elements exhibit moderate to large enrichments relative to primitive earth mantle abundances (Fig. 7a). Sr and Ba have variable degrees of enrichment ranging from .3 to 9 and 4 to 400, respectively. The rocks are LREE enriched and have large to moderate negative to slightly positive Eu anomalies (Fig. 7b).

Isotopic compositions of the rocks of the Agamenticus Complex are few. A  $\text{Sr}^{87/86}$  initial ratio of 0.7108 was calculated by Hoefs (1967). In this study, a  $227 \pm 3$  Ma whole rock isochron was generated using all phases of the complex.

## DISCUSSION

Understanding the petrogenetic evolution of the Agamenticus Complex is a multifaceted problem. The most basic issue, and the one that is constrained the worst, concerns the nature of the primary melt that gave rise to the exposed rocks of the complex. Two models commonly used to explain the petrogenesis of felsic alkalic or A-type felsic rocks involve fractional crystallization of an alkali olivine basalt derived from partial melting of the mantle (Nelson et al., 1987; Loiselle, 1978) or partial melting of a dehydrated granulitic crust (Collins et al., 1982). The realization that rifting is accompanied by vast amounts of basalt production in the mantle and that this greater influx of heat into the crust can produce partial melts of the lower crust (Hildreth, 1981) provides credence to each of these hypothesis. Another petrogenetic concern is to account for the variety of rocks presently seen in the complex. Fractional crystallization accounts for most geochemical trends observed within other suites of rocks with similar petrologic affinity to the Agamenticus Complex, however, crustal assimilation and magma mixing are suggested to explain various geochemical perturbations (Loiselle, 1978; Nelson et al., 1987; Barker et al., 1975; Czamanski et al., 1977). Finally, it is important to understand the post-magmatic history of the complex, specifically the role of deuteric alteration. Many mineralogic and petrochemical changes may occur that confuse the magmatic signature.

The most primitive magmas that occur in the Agamenticus Complex are the mafic syenites and the syenite within the SQSZ. Studies of rocks with similar compositions, for example, the Trans-Pecos trachytes, suggest that their origin can be modeled by fractional crystallization of an alkali olivine basalt using fosteritic olivine, augite, calcic plagioclase and Fe-Ti oxide as the fractionating assemblage (Nelson et al., 1987). With respect to the Agamenticus Complex, this petrogenetic hypothesis is circumstantially supported by the occurrence of the syenites at the evolved end of the basalt to syenite evolution trend on the agpaitic compositional diagram (Fig. 6). Increasing agpaitic index as observed in this diagram is usually associated with increased amounts of crystal fractionation of alkalic rocks. However, unlike many of the WMS bodies in New Hampshire, the Agamenticus Complex lacks observable basaltic rocks and high amplitude magnetic and gravity anomalies that indicate the presence of a near surface basaltic component. Unless significant amounts of a basaltic parent are located in the deep crust, it is difficult to envisage the evolution of the felsic rocks of the Agamenticus Complex by fractional crystallization of a basaltic magma.

An additional concern with this hypothesis is the high  $\text{Sr}^{87/86}$  initial ratio of .7108 (Hoefs, 1967). Unless this ratio is incorrect, fractional crystallization of an alkali olivine basalt must have been accompanied by an unreasonably large amount of assimilation of quite radiogenic crustal material. Alternatively, the high  $\text{Sr}^{87/86}$  initial ratio suggests that partial melting of old radiogenic crust is a viable process. A process involving multi-stage



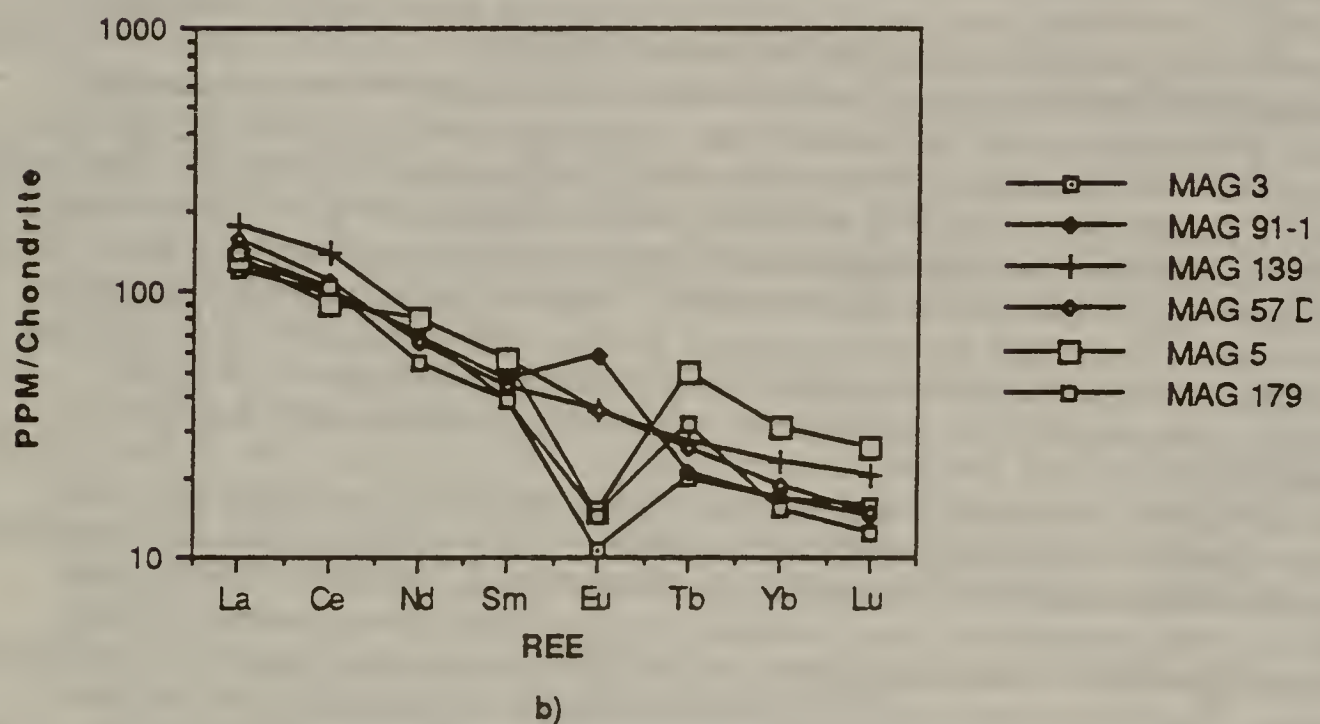
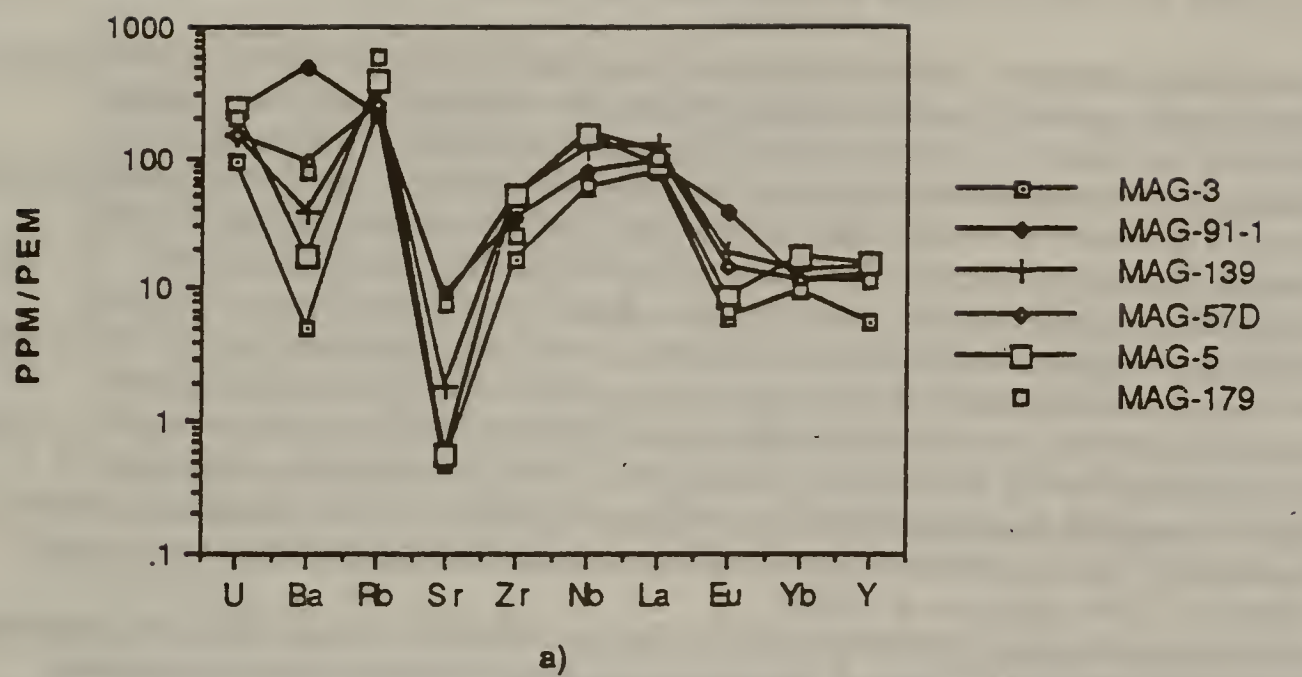


Fig. 7 - Normalized values of selected trace and rare earth abundances. a) Normalized to primitive earth mantle (Taylor and McLennan, 1981) b) REE normalized to chondritic values of Haskin et al. (1968).

melting of the crust has been proposed for the Pikes Peak Complex by Barker and others (1977).

The second petrogenetic question that must be addressed involves the relationship of the syenites, alkalic granite(s) and the porphyritic biotite-amphibole granite. Relative ages of these units indicate that, at least in terms of emplacement, the syenites are first, followed quickly by the alkalic granite(s), with the porphyritic biotite-amphibole granite last. Geochemical similarities between the syenites and the alkalic granite(s) indicate that these magmas are probably cogenetic. The order of crystallization observed for the syenites, plagioclase + alkali feldspar, hedenburgite, fayalite, and amphibole, constrains the fractionating assemblage from the syenites to produce the alkalic granite(s). The appearance of alkali feldspar as the liquidus phase and a lessening of the importance of plagioclase as a fractionating phase could explain the observed changes in the trends of  $K_2O$ ,  $Al_2O_3$ , Eu, and Sr (Fig. 4) (see also Nelson et al., 1987). The relative depletion of Sr and Ba and the large Eu anomalies (Fig. 7) are also compatible with the fractionation of plagioclase and alkali feldspar from the more mafic syenites to the alkalic syenite and alkalic granite(s) (Fowler, 1988; Buma et al., 1971).

The porphyritic biotite-amphibole granite may have evolved independently from the syenites and alkalic granites. On many of the major and trace element plots the porphyritic biotite-amphibole granite is separate from the rest of the complex (eg.  $Fe_2O_3$ ,  $P_2O_5$ , Sr; Fig. 4). The granite also plots in a different field on the agpaitic diagram (Fig. 6) and in discrimination diagrams such as Rb vs Y+Nb (Pearce et al., 1984) and R1-R2 (Batchelor and Bowden, 1985). It is likely that the porphyritic biotite-amphibole granite resulted from crustal melting. This melting may owe its origin to the passage and occasional stagnation of other magmas through mid-crustal levels (Barker et al., 1977).

Several petrographic and geochemical features of the Agamenticus Complex are incompatible with simple crystal fractionation and suggest that other processes played a part in the petrogenesis of the complex. Two trends can be distinguished for rocks with less than 65 weight percent  $SiO_2$  on  $Al_2O_3$  and  $Fe_2O_3$  Harker diagrams (Fig. 4). These trends separate the mafic syenites and amphibole-bearing alkalic syenites (trend A) from the aenigmatite-bearing syenites and the SQSZ (trend B). Trend A may represent a mixing line between the mafic syenites and the non-amphibole-bearing alkalic syenites. Support for this idea is twofold; 1) olivines with reaction coronas of pyroxene occur in the amphibole-bearing alkalic syenites and 2) the occurrence of mafic syenite within the amphibole-bearing alkalic syenites.

On the other hand, trend B which includes rocks of the SQSZ, may signal metasomatic alteration by post-magmatic deuteric activity. Many of the textures and mineral chemistries observed within rocks of the SQSZ indicate deuteric alteration. This process would alter the rock compositions in a systematic, yet highly variable manner. It is likely that the fluids responsible for this alteration in the SQSZ were associated with the intruding alkalic granite. The wide scatter in Rb, U, Th and decoupling of geochemically similar elements (eg. Rb and Ba) provide geochemical support for the interaction of fluids. An alternative explanation of these trends involves the mixing of magmas with different compositions. In their study of the Kaerveen Complex, Greenland, Holm and Praegel (1988) suggest that similar geochemical variations, as well as a strong Sr-Ba correlation (also seen in the Agamenticus Complex) may indicate magma mixing of syenite and alkalic granite.

Many unanswered questions remain concerning the petrogenesis of the Agamenticus Complex. Continued field mapping, additional phase and rock chemistry, and isotopic studies are underway to help unravel some of the mysteries.



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## ITINERARY

**ASSEMBLY POINT:** Meet in the rest area/ official information on the north bound side of I-95 3.1 miles north of Piscataqua River Bridge. Park in the rear behind the information building.

### Mileage

Total      Elapsed      Comments

0		Rest Area
3.3	3.3	Return to I-95N to Exit 1, "To Rts. 1 and 91, York, Ogunquit"
3.75	0.45	Left at end of exit ramp. Cross I-95.
4	0.25	Right on Chases Pond Road. straight at first Y intersection.
6.45	2.45	Intersection with Scituate Road; go straight
7.7	1.25	Continue straight onto Mountain Road. (Mt Agamenticus Road on topographic sheets).
10.5	2.8	Right onto tarred road; go to top of Mt Agamenticus.
11.1	0.6	Top of Mt Agamenticus. Park at end of parking area to right of fire tower.

**Stop 1:** Pavement outcrops of hypersolvus, quartz syenite to alkaline granite are in this locality. Alkaline granite is best exposed on the right hand edge of parking area. Outcrops of quartz syenite to alkaline granite with harrasitic layering of amphibole and xenoliths of Kittery Fm. and mafic hornfels can be found at the top of old ski trail head beyond the parking lot. Alkaline granite to the west of Mt. Agamenticus contains abundant metasedimentary xenoliths possibly indicating a close proximity to the roof of the magma chamber. The alkaline granite is dominated by medium-grained, euhedral to subhedral, patch and string perthite and subhedral quartz. Prior to exsolution of nearly pure albite and orthoclase lamellae, the alkali feldspar composition was Or<sub>50-55</sub>. Interstitial microcline, albite, quartz and amphibole are present. Granular plagioclase rims most perthite grains.

		Return to the bottom of mountain access road.
11.85	0.75	Right onto Mountain Road.
12	0.15	Top of hill pull off to left side of road. Outcrops on either side of the road.

**Stop 2:** Quartz syenite to alkaline granite (Mag 5, Table 1 and 2) similar to that observed at Stop 1 crops out on either side of the road. Farther west, the alkaline granite is more quartz rich (Mag 23+26, Table 2). Amphibole is included in the rims of and is interstitial to the perthite. Amphibole (MAG 193, Fig. 3) ranges from barrosite to arfvedsonite (Following Giret et al., 1980). Arfvedsonite is interpreted as the product of late deuteric growth. The turbid appearance and complete unmixing of the perthite, the presence of interstitial microcline, and the replacement textures of the amphiboles observed in thin sections provide evidence for a late deuteric event.





**Stop 5:** This stop examines a series of outcrops that range from coarse-grained alkaline syenite to fine-grained alkaline syenite and quartz syenite. This variation is tentatively interpreted as a contact zone which has been truncated by the alkaline granite and/or the mesocratic syenite to be seen at Stop 9.

All of the outcrops are located along the power line which crosses Mountain Road. Stop 5a is to the north of the road on the other side of a low lying wet area. The rock at this outcrop is a dark green, coarse-grained alkaline syenite, typical of the southern portions of the alkaline syenite (Fig. 2). It consists dominantly of euhedral to subhedral perthite with interstitial fayalite and hedenbergite (MAG 3, Table 1). In the rest of the syenite body barrositic amphibole forms a reaction rim around the hedenbergite (MAG 61b, Fig. 3). Small alkaline granite dikes are located on the eastern portion of this outcrop and are interpreted to be stringers from the eastern lobe of the alkaline granite.

Stop 5b is located on the southern side of Mountain Road (MAG 2, Table 1 and 2). This medium-grained syenite contains the same mineralogy as Stop 5a. Clots of alkali feldspar can be seen on the weathered surface. These clots are interpreted to be cumulate and indicate that alkali feldspar fractionation has played an important role in the petrogenesis of the complex.

As you proceed over the hill to the south of the road, the syenite locally becomes finer grained and more quartz rich. Fresh samples can be observed at the base of the power line poles. At pole #64 several of the blocks brought up by blasting have a mottled appearance that is typical of portions of the mesocratic syenite to the west. The medium to fine-grained syenite to quartz syenite is continuous to the south up to pole #69 (We'll stop at pole #61 or 62 today). At this locality quartz, feldspar, amphibole stringers occur within the outcrop. These are similar to the amphibole concentrations that occur at the contact of the alkaline granite elsewhere. This contact relationship will be seen at many of the following stops. Within several tens of meters to the south of pole #96 alkaline granite becomes the dominant lithology. These relationships indicate that alkaline granite may underlie the syenite at Pole #69 along a subhorizontal contact.

		Return to the car along power line and then continue eastward on Mountain Road.
16.5	0.35	Right on Rt. 1
16.8	0.3	Left onto River Road. in Town of Cape Neddick. Those who did not bring a lunch should stop at one of the stores at this intersection.
17	0.2	Left into Cape Neddick Baptist Church parking lot.

**Stop 6:** Outcrop on left of parking lot is typical of the southeastern portion of the "contaminant zone" or quartz syenite zone as mapped by Hussey (1962). This outcrop contains porphyritic, medium to dark gray syenite intruded by an alkaline granite to quartz syenite. Amphibole concentrations are common within restricted portions of intrusive fingers and suggest an increase in the volatile content in these regions.

		Continue east on River Road.
17.45	0.45	Right at end of road.
17.8	0.35	Take right at stop sign onto Shore Road. And park along side of road between here and Cape Neddick Campground which is located on other side of small bridge.
18.1	0.3	Cape Neddick Campground



**Stop 7: Lunch Stop.** Permission must be obtained from the owners of the campground to visit this stop. This stop demonstrates the intrusive relationships of the eastern lobe of the alkaline granite (MAG 48 and 49, Table 2) with the Kittery Formation. As seen at Stop 6, amphiboles are often concentrated within the more restricted small dikes or fingers of alkaline granite. Also present is a late cross-cutting basalt dike which are common within both the Agamenticus Complex and the country rock within the coastal region (Hussey, 1962; Swansen, 1982). The relationship of these dikes (if any) to the Agamenticus Complex is unknown.

		Continue south on Shore Road.
18.4	0.3	At stop sign go straight onto Rt. 1a S.
19	0.6	Bear right at Goldenrod Kisses in York Beach. No stop today, but a sharp left will lead along a rocky shore where the eastern lobe of the alkaline granite is well exposed. Access is limited so park in town and enter from beach.
19.15	0.15	Sharp right just as leaving town onto Bridge Road.
19.6	0.45	Right onto Rogers Road. Follow to intersection with Rt 1.
20.5	0.9	Intersection with Rt. 1. Take right and then almost immediate left. Rt. 1 is a busy highway so make sure both northbound lanes are empty before turning.
20.55	0.05	Left into parking lot of United Methodist Church. Park next to Rt. 1 across from brass bell. Outcrops on Rt 1. Narrow shoulder so be careful of traffic!

**Stop 8:** This stop provides a better view of the intrusive relationships observed at Stop 6. Alkaline granite to quartz syenite brittely intrudes a dark gray, porphyritic, aenigmatite-bearing syenite (MAG 139, Table 1 and 2). The power line that crosses Rt. 1 just to the north of this outcrop is the same one that we walked along at Stop 5. Approximately 1 km. north on this power line, the contact between the porphyritic syenite and the alkaline granite can be observed at pole #75. The porphyritic syenite is not observed to the north of this contact along the powerline.

		Continue south on Rt 1.
20.85-21.05		No stop. Outcrops of dominantly alkaline granite with xenoliths of porphyritic syenite
21.4		Contact of alkaline granite and Kittery in valley before Faircrest Motel.
21.55		Kittery Fm in small quarry on west side of road.
22.85	2.3	Take right at lights onto connector road for I-95.
23.35	0.5	Right onto Chases Pond Road
24.0	0.65	Left onto Scituate Road.
24.5	0.5	Scituate Road bears sharply to right.
25.25	0.7	Outcrop located on right side of road. Go past and turn around in driveway on left and park on same side of road or along driveway.

**Stop 9:** At this stop the three dominant phases within the "contaminant zone" can be examined. The rock at northern end of the outcrop is a medium-grained, brown to gray weathering, quartz-bearing, mesocratic syenite (MAG 11a, Table 2). This rock has a mottled appearance, in which the early amphiboles appear to be rimmed by potassium feldspar. This petrographic feature is typical of the "contaminant zone" from here to I-95. An aenigmatite-bearing, porphyritic syenite (MAG 11b, Table 1) is located in the central

and southern portions of the outcrop. On the southern end of the outcrop and continuing to the back of the outcrop a xenolith-rich zone of alkaline granite to quartz syenite can be observed (MAG 11c, Table 2). A large variety of compositions and textures can be seen in the angular and rounded xenoliths. In the central portion of the outcrop the xenolith-rich phase intrudes the porphyritic syenite and the mesocratic syenite. Amphibole concentrations occur at the contact of the alkaline granite with the mesocratic syenite. This can be observed both on the pavement and vertical surfaces and is marked by a change in outcrop appearance from a blocky one of the alkaline granite to a rubblely one of the mesocratic syenite.

		Continue south on Scituate Road.
26.1	0.85	Go straight
26.4	0.3	Right onto Fall Mill Road
26.8	0.4	Cross over river. Eliot Fm. occurs in river bed where it is cross cut by late mafic dikes.
26.95	0.15	At stop sign take right onto "Dead End" road.
27.1	0.2	At top of hill Eliot Fm exposed on right
27.25	0.15	Road to next stop is on right. Turn around and park on side of road. A mafic dike is exposed on east side of road. No stop now but a well exposed outcrop of alkaline granite is located farther on dead end road on east side.

Walk down dirt road. Go straight at the fork to bottom of hill. Turn left at bottom of hill. On the left between this corner and next fork, the contact between alkaline granite and Eliot Fm is located within small depression. No stop.

Take a sharp right at next fork. Go approximately 60 paces to the top of the hill. Outcrop is located in small riverbed slightly off the road to the left.

**Stop 10:** Several intrusive relationships are seen at this outcrop; 1) the contact of the western lobe of the alkaline granite with the Eliot Fm., 2) a basalt dike along the river bed, 3) remnants of a trachytic dike as a scab along the vertical walls of the riverbed, and 4) the informally named leopard rock.

The contact between the alkaline granite and the Kittery Fm is exposed twice as one walks downstream from the road. The orientation of this contact is N45W with a northward 40 to 60 dip. The orientation of the contact does not appear to be controlled by the orientation of the Eliot Fm. bedding (N2E 42S). The orientation of the contact measured here is similar to a contact measurement obtained on the western end of Boulter Pond. The geometry of the contact with the country rock has not been observed elsewhere along the perimeter of the complex.

Along both contacts of the alkaline granite and Eliot Fm. there is an approximately 1 meter wide zone in which the 'leopard rock' occurs. This rock is comprised of subrounded to rounded dark gray, clinopyroxene-rich mafic clasts or blobs (MAG 125CP, Table 2) contained within the alkaline granite. Two possible explanations for the origin of this rock are: 1) Time separated, multiple intrusion along zones of weakness, 2) Intrusion of two coexisting magmas along the same zone resulting in magma mixing. The composition of the mafic clasts is similar to that of a clinopyroxene-bearing dike that intrudes the syenitic phase located on I-95. Based on cross cutting relationships, that dike is older than the alkaline granite but younger than or coeval with the syenite. The interpretation of the mafic clasts as an earlier phase (hypothesis 1) at this locality is supported by their hornfelsic texture and absence of quench textures expected when mafic and felsic magmas



intermingle. This texture is atypical of that expected from quenching and suggests that the mafic phase had cooled prior to the intrusion of the alkaline granite. Please use your discretion when sampling this rock!

This was last stop. Those who are in their own vehicles, and do not need to return to the rest area, retrace route to I-95 and go north to continue to Farmington. Those needing to return to rest area retrace route and cross over I-95 to Rt 1. Go south on Rt 1 for 6.2 miles. Take right onto access road for rest area. Will need to get to adjoining parking lot by going through small gate (with "Do Not Enter" sign) and turning left. Return to I-95 north to proceed to Farmington. Gas stations are available on Rt 1 south just past first set of lights.

#### ACKNOWLEDGMENTS

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## ILLINOIAN AND LATE WISCONSINAN TILLS IN EASTERN NEW ENGLAND: A TRANSECT FROM NORTHEASTERN MASSACHUSETTS TO WEST-CENTRAL MAINE

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### INTRODUCTION

During the previous NEIGC meeting in New Hampshire, at the first stop of trip B-4 (Ridge, 1988), three superposed diamicton units exposed in a gravel pit prompted discussion by trip participants as to whether the units were representative of the "two tills" of southern New England. An overview of the "two-till problem" can be found in Schafer and Hartshorn (1965), Koteff and Pessl (1985), and in this report. A discussion of the Pleistocene stratigraphy of New England is found in Stone and Borns (1986).

Some of the pertinent questions relative to the "two-till problem" outlined by Ridge (1988) include the following:

- 1) Are the two units the product of a single glaciation or is there evidence of more than one ice advance?
- 2) Is the oxidation on the surface of the lower till recent or a function of weathering prior to the deposition of the overlying units?
- 3) Can lithology be used consistently to identify the upper and lower tills?

Similar questions have been discussed by many others, as early as Upham (1878) (and cf. Koteff and Pessl, 1985, and see this report), however it has been almost 20 years since an NEIGC trip specifically addressing the "two-till problem" and its relation to till stratigraphy elsewhere was offered (Pessl



and Schafer, 1968; Borns and Calkin, 1970; Pessl and Koteff, 1970; Shilts, 1970). Borns and others (1970) discussed problems concerning till stratigraphy in west-central Maine, southeastern Quebec, and northern New Hampshire. Several problems presented in these earlier studies have since been addressed, however, a unified regional correlation of till units has yet to be established.

The ages of the two tills has been a point of contention for as long as the two tills have been described. The tills have been argued to be the result of a single glaciation or of multiple glaciations, and have generally been assigned ages of Late Wisconsinan for the upper till and pre-Late Wisconsinan for the lower till. Work by Oldale and others (1982) has allowed Oldale and Eskenasy (1983), Oldale (1987), and Oldale and Colman (1988) to propose correlation of the Illinoian or older aged till at Sankaty Head with the lower till of the mainland. This older till is pre-oxygen-isotope stage 5 in age, and most likely corresponds to oxygen-isotope stage 6 (Oldale and Colman, 1988). It is this correlation to which the authors refer when the lower till and lower till equivalents we shall see on this trip are described as having an Illinoian age.

Definitive criteria which prove the relative age of the two tills include the following:

- 1) Sites where glacial meltwater deposits are associated with the Illinoian glaciation and are bounded by Illinoian and Late Wisconsinan till, as at Nash Stream, New Hampshire (Koteff and Pessl, 1985);
- 2) Thick weathering zone (5 meters or greater) and degree of mineral alteration in the oxidized lower till as compared to that of oxidized upper till (Schafer and Hartshorn, 1965; Stone, 1974; Newton, 1978; Koteff and Pessl, 1985; Newman and others, 1987);
- 3) Stratal disruption along the contact between the two tills where blocks of oxidized lower till are found within the overlying nonoxidized or weakly oxidized upper till (Pessl, 1966; Pessl and Schafer, 1968; Pease 1970; Koteff and Pessl, 1985; Thompson, 1986; Thompson and Smith, 1988).
- 4) Dating methods; dates on material from interglacial deposits are restricted to sites on Nantucket Island and Long Island. Dates from deposits on the mainland have been either minimum or maximum ages, and have never been from units interbedded between the two tills (Stone and Borns, 1986).

Other criteria which by themselves are not diagnostic of the units, but which have been presented as evidence for age discrimination include the following: 1) till facies, 2) color, 3) hardness, 4) grain size, 5) iron-manganese stain along fractures and coating stones, and 6) local stratigraphic position.

Part of the purpose of this trip is to introduce the problem to uninitiated parties and present examples based on field criteria of the two till units of southern New England, as well as other units farther north which are problematical as to how they fit into the story. This field trip will

focus on field observations and data collected at several locations in the summer of 1989 at sites in northeastern Massachusetts and southwestern Maine (stops 1-6, Figure 1) to aid in the recognition of varieties of and the mappability of the "two tills". The last two stops (7 and 8, Figure 1) will present a site where multiple tills were considered to represent multiple glaciations, but recent work indicates that the tills can be shown to be a result of a single glacial cycle, the Late Wisconsinan.

## TILL STRATIGRAPHY OF SOUTHERN NEW ENGLAND

by Byron D. Stone

### Terminology

At present, tills of southern New England are designated by informal stratigraphic names or by proposed formal names for local varieties of till (Figure 2). No regional study of varietal members of the upper, surface till of late Wisconsinan age or the lower, pre-late Wisconsinan till in drumlins, complete with laboratory analyses, supports an inclusive formal nomenclature for the two tills. In the field, physical criteria differentiate local varieties of the tills, always on a basis of comparative characteristics in areas of similar bedrock type. Because of the obvious close relation of till composition and texture to very local bedrock lithologies (Flint, 1930, Samuels, 1962, Force and Stone, in press), differences within one till unit may be greater in some areas than differences between varieties of both units.

The general terms **upper till** and **lower till** (Schafer and Hartshorn, 1965, Stone and Borns, 1986) seem confusing to some workers because the terms emphasize an expected superposition of the units. In most exposures, this superposed stratigraphy consists of a very thin (<1 m) sandy upper till with cobbles overlying a mixed-till zone that contains discrete angular fragments of the lower till within a rusty-oxidized sandy matrix. Thick, compact gray upper till is present above such a mixed zone or above oxidized lower till in relatively few exposures (Pessl and Schafer, 1968). The upper till forms a discontinuous till blanket of highly variable composition over the upland bedrock landscape and is present as a thin unit beneath most stratified glacial deposits (Figure 3). Its composition likewise reflects the local source materials in drumlins. The lower till is preserved virtually in only drumlins (Figure 1) and related thick bodies of till with glacially smoothed and streamlined morphology. Probably the two tills do not occur as laterally extensive superposed sheet bodies in the region. The terms **new till** and **old till** have also been used (Schafer and Hartshorn, 1965), in reference to the relative and correlated ages of the units and their respective glacial episodes.

Other confusing terminology has been used. Upham (1880) referred to a compact **lower till** in Boston Harbor drumlins (his **drumlin till**, Upham, 1897) and an overlying thin, loose, oxidized **upper till**. Crosby (1890) called the compact, nonoxidized till in drumlin cores **lower till**, overlain by his **upper till**, the oxidized zone at the surface of the drumlins. Detailed studies of the upper, surface till in New Hampshire (Drake 1971), and western Connecticut (Pessl and Schafer, 1968, Newton, 1978, Smith 1984, 1988) differentiated a loose, sandy unit, containing boulders and cobbles and lenses of sorted and stratified sediments, from an underlying compact sandy till. These studies





Figure 1. Map showing drumlins and generalized glacial marine limit in north-east Massachusetts, southeast New Hampshire, and southwest Maine, and field trip stops. Compiled by B.D. Stone (NH, MA), and T.K. Weddle (ME); based on Alden (1924, revised by Stone; solid drumlins), Thompson and Borns (1985a, open drumlins), and B.D. Stone (unpub. data). Marine limit represented by heavy line.





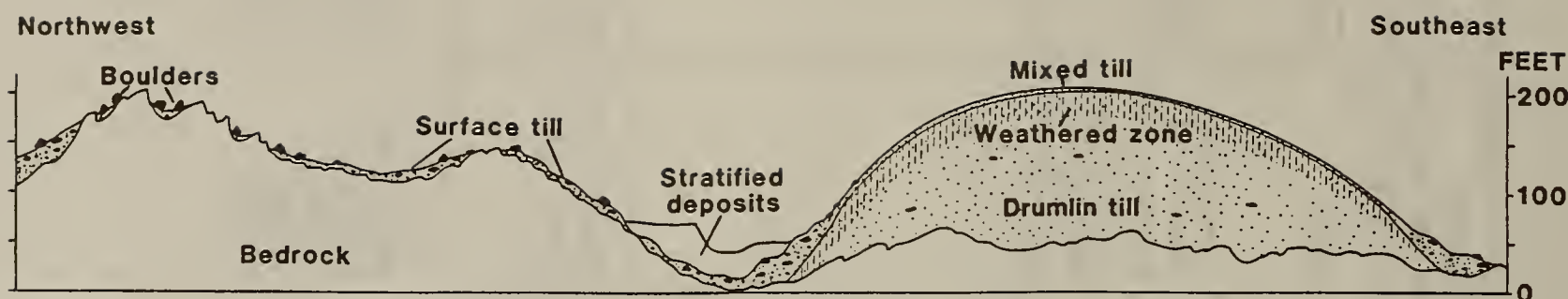


Figure 3. Distribution and relation of surface till (Late Wisconsinan), drumlin till (Illinoian), weathered zone at top of drumlin till, and mixed-till zone, which overlies the weathered zone. Based on numerous exposures, geologic and soil-series map data, and schematic distributions of units shown at field trip Stops 1 and 2.

concluded that the loose till is an ablation facies, comprising debris of probable englacial or supraglacial origin. This dual till-facies classification of the sandy upper till has proven confusing in the past, especially in light of the previous terminology of the two-glaciation genesis of the **upper** and **lower tills** of the region (see discussions in Goldthwait, 1971, and Koteff and Pessl, 1985).

In this section of the report, the **upper till** of southern New England is referred to as the **surface till**. The surface till is the till sheet of Late Wisconsinan age in the area; it is the material referent of the diachronic glacial episode in southern New England from about 24,000 yrs BP to deglaciation about 15,000-14,000 yrs BP. It is dated by radiocarbon dates from preglacial sub till materials incorporated in the drift and from postglacial materials that overlie the drift (Stone and Borns, 1986). The surface till is highly variable in composition, dependent on the composition of local bedrock and older surficial materials. It includes a compact basal unit, of lodgement origin (Stop 1), and a discontinuous, thin overlying unit of loose sandy till, of ablation origin. It also includes the mixed-till zone that overlies the weathered drumlin till in drumlins.

The **lower till** is referred to in this part of the report as the **drumlin till**, in obvious reference to its virtually exclusive distribution in glacially smoothed landforms (Figure 1), which were resistant to Late Wisconsinan glacial erosion. The drumlin till is the locally preserved till of Illinoian age in the region. The drumlin till is the material referent of the diachronous Illinoian glacial episode older than about 133,000 yrs BP. It is dated by consideration of the depth and degree of the weathering in its upper part, and by correlation with the Sankaty lower till which lies beneath dated marine beds of Sangamonian age (Oldale and others, 1982). The weathering zone in the upper part of the drumlin till is related to a

relatively long or intense period of weathering that postdated drumlin formation in the region. Only a compact basal drumlin till unit is known; its texture and composition vary with bedrock composition.

### Previous Work

The tills of New England initially were related to drift of the last glaciation (Upham, 1878, 1879a, 1879b, 1880) and early studies of till in the region (Upham, 1889, 1897) retained a single-glacial framework (Figure 2). Crosby's (1890) pioneering textural analysis (the first in North America) of compact till from drumlins in Boston Harbor was based on the single-glacial premise. Crosby used appropriate descriptive terminology for the deep drumlin-cliff exposures: his "upper till" was the oxidized zone beneath the surface of the drumlin, which overlies nonoxidized gray "lower till" in the drumlin cores. Crosby's original estimate of the volumetric proportion of boulders and stones in the till, and his sieve data are useful today in estimating a whole-sediment grain size distribution (Figure 4). Evidence of multiple glacial stratigraphy in New Jersey (Salisbury, 1892) and New York (Woodworth, 1901) prompted Fuller (1906) and Clapp (1906, 1908) to suggest that compact tills beneath the surface sandy till in Massachusetts and Maine were products of pre-Wisconsin glaciations. Fuller (1906) clearly suggested an Illinoian age for the drumlin till, correlating it with the Montauk Till of the emerging four-fold glacial record on Long Island (Fuller, 1914). This stratigraphy was extended to Martha's Vineyard (Woodworth and Wigglesworth, 1934), and correlated with Wisconsinan and Illinoian glaciations on Block Island and Nantucket. Crosby (1890), LaForge (1932), Alden (1924), Flint (1930), and Denny (1958) described shallow surface exposures of the drumlin till in New England but concluded that the drumlin till and the surface till are both products of a single Wisconsin glaciation.

Crosby (1908) reinterpreted some elements of the Long Island stratigraphy, and suggested that the Montauk and other tills derived from a single Wisconsin glaciation. Similarly, MacClintock and Richards' (1936) review of the regional marine and glacial units of New Jersey and Pennsylvania included their assignment of all tills exposed at the surface of Long Island, including Montauk Till, to Wisconsin age. Results of detailed work in Massachusetts (Currier, 1941, Jahns, 1953) emphasized the deep weathering zone in the upper part of the drumlin till, which indicated a pre-late Wisconsin age of the till. Judson (1949) suggested that drumlin till might be of early Wisconsin age. Flint (1953) adopted the early Wisconsin age of the drumlin till as well as correlation of it with the Montauk. This age assignment was subsequently retained in regional summaries (Muller, 1965, Schafer and Hartshorn, 1965, Sirkin, 1982). Stone and Borns (1986) accepted the early Wisconsinan age of the drumlin till and Montauk Till on the basis of the inferred ages of enveloping marine units on Long Island and on the basis of the revised amino-acid racemization estimated age (Sangamonian) of a shell in the drumlin till from Boston Harbor (Belknap, pers. comm. 1982).

Exposures of oxidized and some nonoxidized drumlin till are known widely across the region from southern Connecticut (Flint, 1961), western Connecticut (Pessl and Schafer 1968, Stone, 1974, Newton 1978, 1979a, Thompson, 1975), eastern Connecticut (White 1947, Pessl, 1966, Stone and Randall, 1977);



western Massachusetts (Newton, 1978), central Massachusetts (Jahns, 1953; Lougee, 1957; Mulholland, 1976; Stone 1980) eastern Massachusetts (Currier, 1941; Moss, 1943; Castle, 1958; Kaye 1961; Oldale, 1962, 1964; Newman and others, 1987), southern New Hampshire (Koteff and Stone, 1971), central New Hampshire (Goldthwait, 1948; Denny, 1958; Newton 1978), northern New Hampshire (Koteff and Pessl, 1985), and southern Maine (Thompson and Borns, 1985b). Figure 3 summarizes the field relationships of the drumlin till and the overlying surface till in these study areas. These relationships may be seen at Stops 1-4 of this field trip.

Detailed studies of the drumlin till included till-fabric analysis in western Connecticut (Pessl, 1971), which demonstrated different fabric directions for the drumlin and overlying surface till. In eastern and central Connecticut, Pessl (1966), Pessl and Schafer (1969), and Pease (1970) noted the locally sharp truncation of the top of the oxidized zone of the lower till by the overlying upper till. They described the mixed-till zone above the truncation: mixed till contains discrete, angular clasts of lower till, both oxidized and nonoxidized, within a sandy upper-till matrix. Koteff and Stone (1971) showed the same relationships and mixed till in southern New Hampshire. Stone (1980) portrayed the distribution of mixed till on drumlins in central Massachusetts, Mickelson and others (1987), and Newman and others (1987) described similar mixed till in Boston Harbor drumlins. Studies of mineralogy and the weathering zone in the drumlin till showed that the upwardly progressive hydration of altered illite in the zone (Stone, 1974) is related to alteration of illite to vermiculite and mixed-layer illite/vermiculite (Newton, 1978, 1979b). Oxidation of iron-bearing minerals, notably garnet, and resultant staining of till matrix (Stone, 1974) also deeply etches surfaces of garnet grains (Newton, 1978). In the chloritic variety of the drumlin till in Boston Harbor, degraded illite is chloritized and altered to a lower charge mineral by oxidation of octahedral  $\text{Fe}^{+++}$  or replacement of  $\text{Fe}^{++}$  by  $\text{Al}^{+++}$  (Quigley and Martin, 1967). This alteration of illite and chlorite to high-charge and low-charge vermiculite also decreases downward through the oxidized zone of the Boston drumlins (Newman and others, 1987).

Materials within the drumlin till are older than the range of  $^{14}\text{C}$  dating ( $>38,000$  BP, Lougee, 1957), but peat in the till at Millbury, Massachusetts contains pollen of interglacial oak, hickory, sweet gum, and minor pine. Shells from drumlins in Boston Harbor include species that are restricted to more southerly, warmer areas (Crosby and Ballard, 1894). Belknap (1979, 1980) calculated an amino-acid estimated ages of 200,000 and 214,000 yrs BP from *Mercenaria* shells in the drumlin till in the harbor. He later recalibrated his paleotemperature model on the basis of a U-series date on coral of Sangamon age from Sankaty Head, Massachusetts (Oldale and others, 1982), and suggested that the Boston Harbor drumlin-till shell is of probable Sangamon age. The maximum age of the drumlin till thus apparently was constrained (Stone and Borns, 1986), supporting the correlation of the till with Montauk Till, previously considered to be early Wisconsinan in age (Sirkin, 1982). On the basis of a regional correlation with the lower till at Sankaty Head, the drumlin till now is inferred to be of Illinoian age.

#### Characteristics of the Surface (Upper) Till and the Drumlin (Lower) Till

Physical characteristics differentiate the two till of southern New England (Table 1). These are related to source materials of the tills, glacial erosional and depositional processes, and weathering effects.

Table 1. CHARACTERISTICS OF THE SURFACE TILL AND DRUMLIN TILL OF SOUTHERN NEW ENGLAND: MEDIUM-COARSE GRAINED IGNEOUS AND METAMORPHIC BEDROCK

	<b>SURFACE TILL</b> (Late Wisconsinan)	<b>DRUMLIN TILL</b> (Illinoian)
Color (naturally moist material Munsell color symbols	Gray to light gray (2.5-5Y 6-7/1-2), to olive gray to light olive gray (5Y 4-6/2)	Olive to olive-gray (5Y 4-5/2=3) to olive-brown (2.5Y 4-5/3-5) in oxidized zone, dark-gray (5Y 3.5-5/1) in nonoxidized till
Texture of matrix ( $<2$ mm [-1 phi], range, fig. 3)	62-80% sand 20-38% silt and clay $<1$ -6% clay	35-60% sand 40-65% silt and clay 11-38% clay
Stone content	19-54% $>2$ mm 5-30% $>76$ mm (3 in)	19-42% $>2$ mm 1-11% $>76$ mm (3 in)
Layering	Textural layering common, generally subhorizontal; consists of thin, lighter sandy layers interbedded	Textural layering not common; thin, oxidized sand layers and vertical sand dikes locally with darker, silty layers; layering is laterally discontinuous
Jointing	None; subhorizontal parting is related to layering and fabric of till matrix	Well developed; closely spaced subhorizontal joints and less numerous subvertical joints impart a blocky or thin platy structure to till
Distribution and thickness	Lies directly on bedrock; less than 3 m thick in areas of rock outcrop; commonly 3-6 m thick on lower valley slopes	Forms cores of drumlins and related bodies of thick till; generally $>10$ m thick, commonly 20-30 m thick; maximum known thickness 70 m
Soils and weathering (representative USDA S.C.S. soil series)	Canton series, Charlton series (Typic Dystrochrepts)	Paxton series (Typic Dystrochrepts); soil developed in mixed-till zone that overlies weathered zone in drumlin till; weathered zone- $<9$ m thick; zone is oxidized, leached in some areas, and contains altered clay minerals and iron-bearing minerals
Geotechnical properties (Unified Soils Classification,	SM, SP-SM, nonplastic	SC-SM, SC, SM, ML PI 10-30, LL $<30$

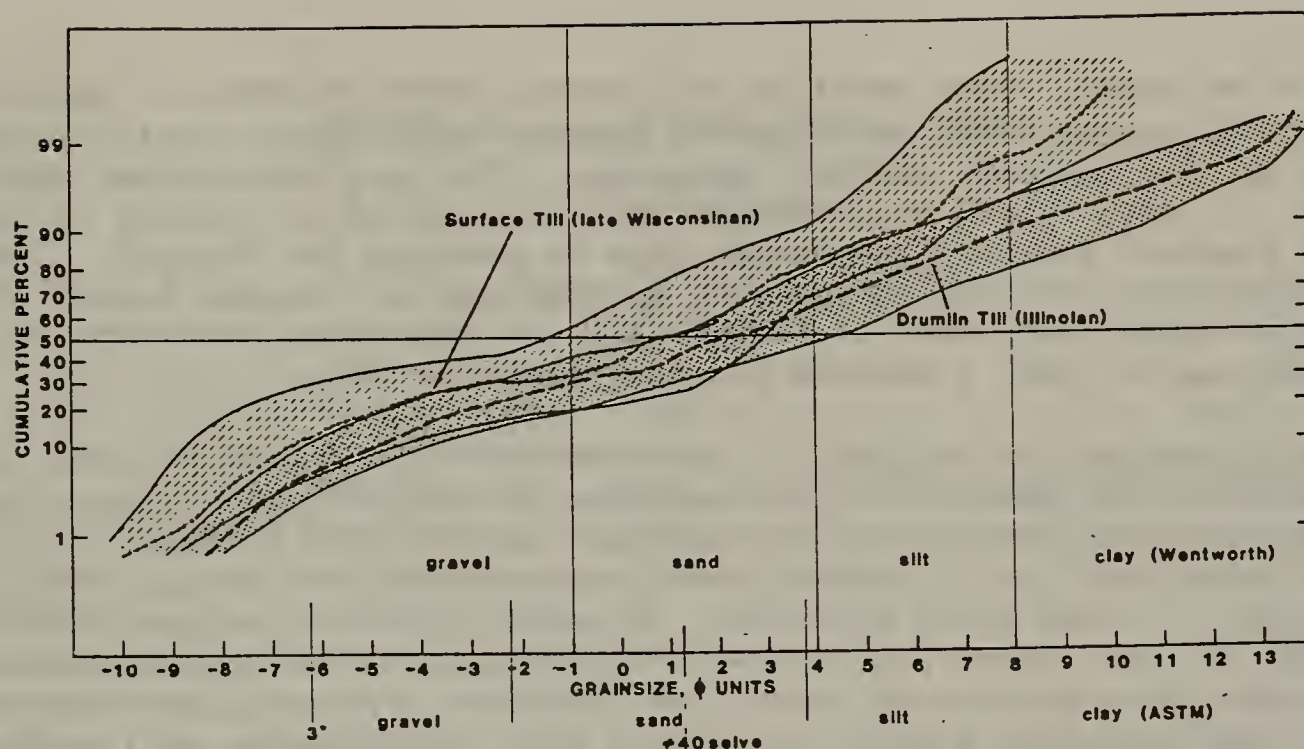


Color. The surface till is generally gray below the present solum, reflecting its composition of fresh minerals and nonoxidized state. Local staining by high chroma Fe+++ minerals, probably limonite, appears to be controlled by ground water movement through materials of contrasting texture or around clasts. The olive color of the weathering zone in the drumlin till is a pervasive oxidation stain (limonite?) that affects all areas of the silty till matrix. It extends through the zone of leaching in some exposures. The stain is darker around iron-bearing minerals. Dark Fe-Mn staining is on joint surfaces and around stones, but generally does not extend as deep as does the pervasive iron stain.

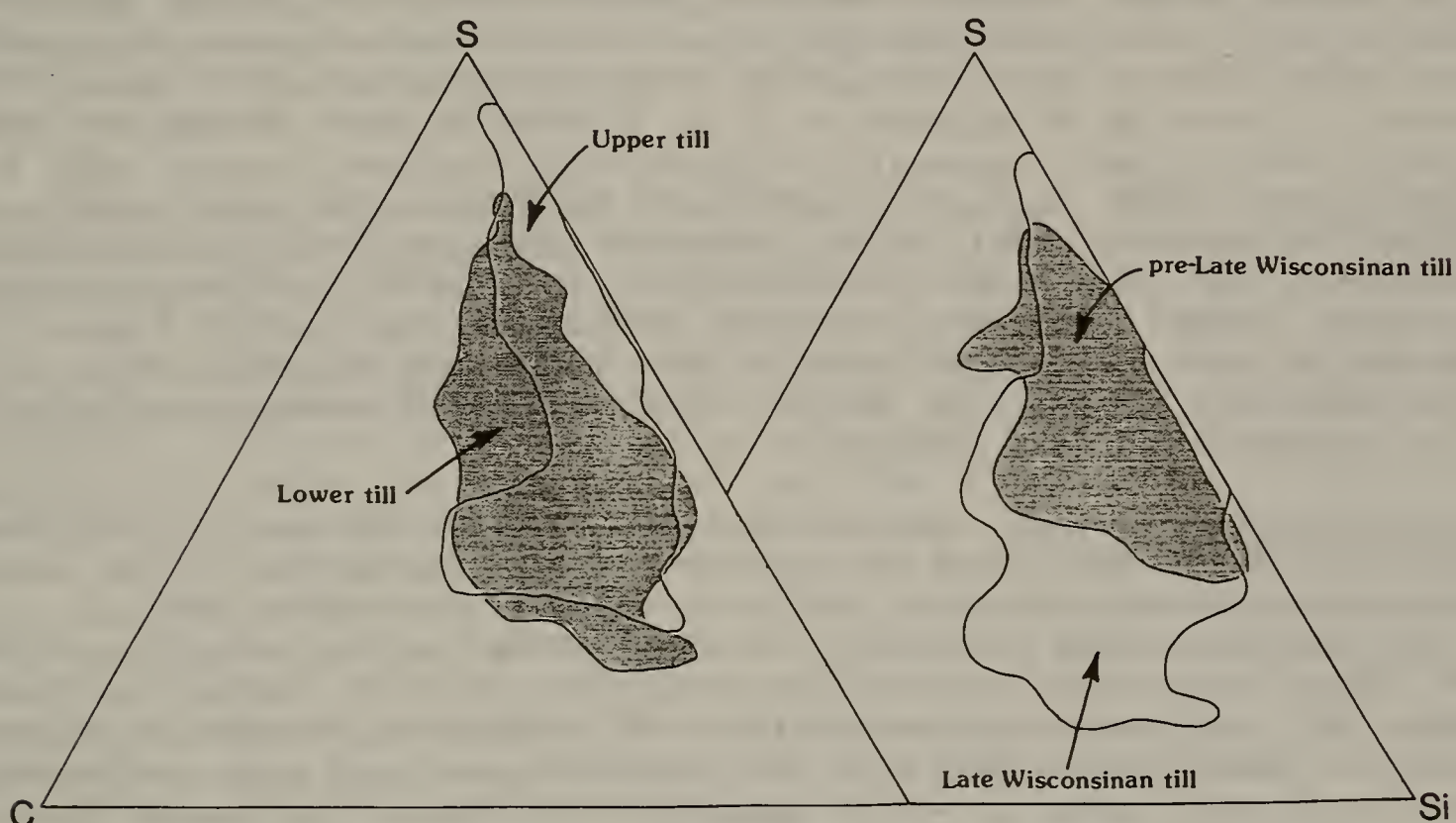
Texture and stone content. Particle-size analysis of the surface and drumlin tills show that whole-till samples differ in stoniness, proportion of the dominant sand-sized particles, and silt and clay content. A consensus has emerged following Crosby's (1890) discussion that the volumetric content of stones larger than two inches in drumlin till is less than 10%, probably about 5% (Pessl and Schafer, 1968, range of values in Fuller and Hotz, 1981). Boulders longer than 1 m are notably rare in large excavations of drumlin till. Accordingly, whole-till particle-size curves are constrained by these visual estimates and gravel sieve data, adjusted for rock/till matrix density contrasts (Figure 4). Likewise, the same workers estimate the stoniness, including large boulders, of the compact surface till to be 5-30% by weight. Boulders 1-2 m in length are common in large excavations of compact surface till (Stop 1); these and smaller boulders are common in the ablation material at the surface of the late Wisconsinan till. The grain-size curve for the compact surface till is drawn to include these visual estimates and sieved gravel intervals. The proportion of sand grain sizes varies greatly within each of the two till units (Figure 4), locally more than samples between units. This variation is related to glacially eroded and comminuted fresh rock fragments and mineral grains that vary with the composition of very local bedrock units (Smith, 1984, 1989; Force and Stone, in press). Fields of sand populations of the two till units overlap (Figure 4). The silt and clay contents of the till matrices are distinguishing characteristics of the two tills. Although the ranges of the distributions overlap (Figure 4), representative values do not. Extreme values show that the surface till contains 20-38% silt and clay, whereas the drumlin till contains 40-65%. The proportion of clay likewise distinguishes the tills: surface till contains <1-6% clay, drumlin till contains 11-38%. The drumlin till also contains a measurable amount of very fine clay (<0.2 microns, about 12.4 phi). Some varieties of surface till may contain clay particle of this size in some basins (varves contain very fine clay), but it generally is not a measurable component.

Distribution and thickness. The late Wisconsinan surface till forms an irregular blanket over bedrock uplands and beneath stratified glacial deposits. In areas of bedrock outcrop the topography of the till surface is controlled by bedrock-surface relief (Figure 3). Here, the till is discontinuous, probably averaging less than 2 m in thickness; it contains numerous boulders. In other areas on northerly facing lower valley slopes, the till forms smooth-to-bumpy patches of true ground moraine. In these areas, the till is 3-6 m thick; the compact basal facies of till forms the





a)



b)

Figure 4. Grain size characteristics of the surface till and drumlin till. a) Grain size distributions; range of grain size distributions and representative ("average") value of drumlin till (dashed) based on data from Crosby (1980), Pessl and Schafer (1968), Stone (1974), Mulholland (1976), Newton (1978), Fuller and Hotz (1981), Koteff and Pessl (1985); values for coarse gravel-to-boulder fraction based on volume estimates of clasts, adjusted for densities of rock and till matrix. Range of grain size distributions and representative value of surface till (dotted) based on data from above authors, except Crosby, and including data of Drake (1971). b) Ternary diagrams of sand-silt-clay values of surface (upper) till and drumlin (lower) till of southern New England and Maine. Fields of values for southern New England based on data from Pessl and Schafer (1968), Stone (1974) Thompson (1975), Mulholland (1976), Newton (1978), Smith (1984), Koteff and Pessl (1985), Newman and others (1987), W. H. Newman (pers. comm., 1989; for Maine, Borns and Calkin (1977), Ackert (1982), Brady (1982), Kite (1983), Newman and others (1985), Lowell and Kite (1986), Weddle and Retelle (1988), Weddle (unpub. data). Upper till  $n = 145$ , lower till  $n = 204$ ; Late Wisconsinan till  $n = 118$ , pre-Late Wisconsinan till  $n = 40$ .



bulk of the unit. Loose ablation till with surface boulders is generally thin in bedrock outcrop areas and areas of thicker till. Only locally is it thick enough to form hummocky surface topography. The late Wisconsinan surface till is the till unit mapped on surficial geologic maps of all states in the region and on numerous detailed quadrangle maps in southern New England. The drumlin till is preserved almost exclusively in drumlins and related bodies of glacially smoothed thick till (Figure 3). It is generally 10-30 m thick in these bodies, and has a maximum known thickness of 70 m.

Weathering and soil development. Soils developed in the upper 0.6-1.2 m of the surface till, mixed-till zone and drumlin till (Figure 3) since Late Wisconsinan deglaciation are inceptisols, characterized by poorly developed cambic B-horizons (<1.2x clay of overlying horizons) and weakly modified clay mineralogy. In the field trip area, representative soil series on the surface till are Canton and Charlton soils in Massachusetts, and the Paxton series on the mixed-till zone on drumlins (Fuller and Hotz, 1981; Fuller and Francis, 1984). The weathered zone at the top of nearly all drumlin till exposures is 3 m to as much as 9 m thick. The base of the oxidized zone is subparallel to the surface of the landform, indicating pedogenesis after glacial smoothing and prior to Late Wisconsinan glaciation. Field observations show progressive weathering effects upward through the zone: pH values decline (Stone, 1974), amount of leaching is progressive at the base of the zone (Crosby and Ballard, 1894), color values of pervasive oxidized stain increase (Crosby, 1890, Pessl and Schafer, 1968), degree and darkness of Fe-Mn stain on joint faces increase (Pessl and Schafer, 1968), blocky structure increases and is more densely developed. Lab data showing alteration of clay minerals and iron-bearing minerals further define the weathering gradient through the 3-9 m zone. The weathering zone is the upper part of the C-horizon of a probable well developed soil (Stone, 1974, Newton, 1978), the solum of which was removed by late Wisconsinan glacial erosion.

Geotechnical properties. Geotechnical properties of the two till units are dependent of the grain-size and water-transmitting properties of the units, which yield characteristically different values. The Unified Soils Classification of the surface till is **SM or SP-SM** (sand>gravel, fines <12%, SM, fines 5-12%, SP-SM; fines of low plasticity index, PI, table 1). These soils are **silty sand with gravel with 3-17% cobbles and boulders by volume** (SM), or **poorly graded sand with silt and with gravel and with 3-17% cobbles and boulders by volume** (SP-SM). Reported variations in the drumlin till have lead to classifications as **SC, SM, SC-SM, ML** (sand>gravel, fines >12%, SC, SM, SC-SM; fines>50%, ML). These soils are **clayey sand with gravel, with 1-15% cobbles and boulders by volume** (SC, PI 10-30, LL <30), **silty sand with gravel, with 1-15% cobbles and boulders by volume** (SM, low PI, low liquid limit, LL), a combined unit (SC-SM), or a **sandy silt with gravel and with 1-15% cobbles and boulders by volume**, ML. Atterberg values (derived from tests of <40-mesh sieve, 1 1/4 , fig. 3) of nonplasticity and LL=0-<10 for the surface till contrast with PI=10-30 and LL=<30 for the drumlin till. In the field, naturally moist samples of the surface till exhibit low dry strength (a measure of "compaction" of fragments): fragments crumble or "pop" with some finger pressure. Drumlin till has medium dry strength: considerable finger pressure is required to pop fragments.



## Southwestern and West-Central Maine

### Previous Work

In Maine, a two-till stratigraphy was described by the early workers, although agreement on whether the tills represent multiple glaciation was never established (Holmes and Hitchcock, 1861; Stone, 1899; Clapp, 1906, 1908; Leavitt and Perkins, 1935). After Caldwell's discovery of the wood in a deposit between two tills at New Sharon (1959, 1960), no other work on multiple glaciation in Maine was published, with the exception of Borns and Calkin (1970, 1977). New Sharon was known by Leavitt and Perkins to be an organic-bearing locality (pers. comm., H. W. Borns, Jr., 1986), however they do not mention it in their Bulletin No. 30 (Leavitt and Perkins, 1935). Quaternary studies in southwestern and west-central Maine over the last two decades generally have not addressed the "two-till problem" and multiple glaciation. The exception to this has been the reconnaissance mapping in southwestern Maine by Woodrow Thompson of the Maine Geological Survey, discussed in Thompson and Borns (1985), Thompson (1986), Thompson and Smith (1988), and abstracts and field guides by Caldwell and Pratt (1983), Caldwell and Weddle (1983), Weddle and Caldwell (1984, 1986), Weddle (1985, 1986, 1988), Weddle and Retelle (1988).

The only geologic quadrangle published by the U. S. Geological Survey exclusively dedicated to surficial geology in Maine is the Poland 15-minute quadrangle (Hanley, 1959). Two distinctive till types are differentiated on this map, a silty till overlain by a sandy till, and where exposed and in superposition the contact is described as gradational between the units. The silty till is described as dusky yellow to dark olive brown, moderately indurated, and tough, and is interpreted as a basal till. It comprises elongate till ridges and drumlins, the long axes of which trend S15-25E. The sandy till is described as grayish yellow, friable, and easily excavated. In places it has a rusty color, presumably derived from limonite staining deposited from circulating ground water. The sandy till contains water-sorted concentrations of gravel, and in places it appears to grade laterally into stratified drift. The distribution of the sandy till and the silty till in the quadrangle is closely associated with the underlying bedrock. The sandy till is generally found in areas underlain by granite of the Sebago Pluton, whereas the silty till is found in areas underlain by metapelite, metadolostone, and metasandstone of the Sangerville Formation (Osberg and others, 1985). While two distinct till lithologies are present in the Poland quadrangle, one of which in places is superposed on the other, Hanley (1959) interpreted them as being derived from different source rocks and not from multiple glacial advances. Stop 5 of this field trip is just beyond the eastern border of the Poland 15-minute quadrangle. At this stop, a sandy diamicton overlies a silty diamicton. However, while their physical characteristics are not quite the same as the general description of the superposed diamictons described in the Poland quadrangle, a lively discussion of their origin is anticipated at Stop 5.



## Comparison of General Characteristics of Till in Southern New England with Southwestern Maine

Color. Similar to southern New England, color has been used by some workers as a criterion to differentiate tills, specifically by the nonoxidized or oxidized state of the till. It is of interest to note that the glacial marine Presumpscot Formation has an oxidized horizon at its surface, extent downward for generally not more than 2 m. This oxidized horizon was thought to represent a different unit from the underlying blue-gray mud, both deposited by different glaciation (Trefethen and others, 1947). Levitt and Perkins (1935), Goldthwait (1949, 1951), Caldwell (1959), and Bloom (1960, 1963) attributed the color difference to post-glacial oxidation of the Presumpscot Formation, and this interpretation has proven correct.

Texture. The general textural characteristics noted in the upper and lower tills of southern New England (upper is coarse grained; lower is fine grained) also have been noted as general field characteristics in similar units in southwestern Maine (Thompson and Borns, 1985). Figure 4 is a compilation of textural analyses from numerous studies in Maine where Late Wisconsinan till and pre-Late Wisconsinan till (southern New England upper and lower till equivalents, respectively) are plotted. While there are two fields shown, there is considerable overlap between the Late Wisconsinan and pre-Late Wisconsinan tills. Of interest is the area in the lower part of the Late Wisconsinan till field. These data from Brady (1982) are predominantly from tills in a region underlain by fine-grained metapelite of the Vassalboro Formation (Osberg and others, 1985). These data from Maine clearly reflect the importance of substrate as a significant control on till lithology. In southern New England, Smith (1984) showed similar findings in till underlain by areas of different bedrock lithologies. However, he also showed that Late Wisconsinan till and Illinoian till differentiated by field characteristics were also distinguishable by texture in areas of different bedrock terrane.

Clay mineralogy. Whereas the clay mineralogy and other diagnostic characteristics have established the presence of a Late Wisconsinan (upper) till and an Illinoian (lower) till in southern New England (Stone 1974; Newton, 1978, 1979a; Newman and others, 1987; W. A. Newman, pers. commun., 1989), there have been no similarly detailed clay mineral studies on any equivalent "lower till" units in southwestern or west-central Maine.

However, Gagnon (Senior honors thesis in geology, Bates College, 1984) examined the clay mineralogy of diamicton samples from test borings at the Hatch Hill landfill in Augusta, Maine. The boring samples were described as having penetrated two tills, an upper olive till and a lower gray till. However, it was never clear if the borings actually penetrated two different aged units or only one unit with a deeply oxidized surface. The x-ray diffractograms indicate severe alteration of the upper unit (illite to mixed-layer clay minerals) with decreasing alteration with depth. There is no indication in the alteration pattern of a less severely weathered unit overlying a more severely weathered unit (that is, a younger "upper" till over an older "lower" till). Because of the nature of the alteration, it may be that the boring samples are from a "lower till" equivalent, however, the report is not detailed enough to determine this.



Weddle (in press, and in prep.) has examined the clay mineralogy of the diamictons at New Sharon and also of the so-called paleosol on the New Sharon Beds. In all cases, the degree of weathering indicated by the clay mineral alteration is not representative of the degree of weathering found in the oxidized "lower till" of southern New England as shown by Stone (1974) and Newton (1978, 1979a).

Soils association. The soils associations presented as characteristic of soils forming on the upper till and lower till of southern New England have not been similarly categorized for tills in Maine, however, the soils found on the drumlins which we will visit in Maine have been mapped (Flewelling and Lisante, 1982). These soils (at Stops 3 and 4) are mapped as part of the Skerry-Brayton-Becket association and the Marlow-Brayton-Peru association. These series of soils are all formed in compact till commonly on the tops and sides of drumlins. The Becket, Marlow, and Peru series are taxadjunct because they have slightly less extractable iron and aluminum than is defined in the range for the series.

#### New Sharon and Regional Correlations

The sections at New Sharon originally described by Caldwell (1959, 1960) have been correlated by others with the southern New England till stratigraphy (Koteff and Pessl, 1985; Thompson and Borns, 1985b), and with the stratigraphy of McDonald and Shilts (1971) of southeastern Quebec (Borns and Calkin, 1977; Stone and Borns, 1986). Recent work at New Sharon has shown that all the exposures there can be attributed to a single Late Wisconsinan event without calling for Early or Middle Wisconsinan multiple glaciations, and that no Illinoian till is exposed at the surface (Weddle and Caldwell, 1984; Weddle, 1986, 1988; Weddle, in press, and in prep.).

The reassignment of the stratigraphy at New Sharon is important because it brings into question the correlations and age assignments by Borns and Calkin (1977), Koteff and Pessl (1985), Thompson and Borns (1985b), and Stone and Borns (1986) of other multiple-till sections in Maine, in particular the sections at Austin Stream and Nash Stream, Maine, and Nash Stream in northern New Hampshire. Similarly, the lack of a paleosol at New Sharon and the allochthonous nature of the New Sharon Beds (Weddle, in press) precludes correlation of the "lower till" of southern New England with any exposed units at New Sharon. The lack of paleosols or the presence of extensive stratified drift interbedded with the diamictons suggests that the entire section exposed at New Sharon was deposited during a single glacial cycle, the Late Wisconsinan glaciation. Currently, the age assignments by Stone and Borns (1986) of different tills in Maine is disputed, and re-evaluation of southwestern Maine and west-central Maine terrestrial Pleistocene stratigraphy is justified.

Furthermore, Fulton and others (1984) assignment of a Sangamonian age (75,000 to 128,000 yrBP; oxygen-isotope stage 5) to the Johnville till and Becancour till in Quebec, and Thompson and Borns (1985b) tentative correlation of the New Sharon till with these units is herein disregarded. An Illinoian age has been proposed for the tills in Quebec (Lamothe, 1987; LaSalle, 1989), and this age is in agreement with the Illinoian age assignment of the lower till at Sankaty Head, Nantucket (Oldale and others, 1982; Oldale and Eskenasy, 1983; Oldale, 1987; Oldale and Colman, 1988), which has been correlated with the lower till on the mainland of southern New England.



We hope that this field trip will provide an understanding of some of the problems of Pleistocene terrestrial stratigraphy in the region, and allow a means of evaluating the correlation of the southern New England till stratigraphy with southern Maine.

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## FIELD TRIP ROAD LOG

**Assembly Point:** Westgate Shopping Center K-Mart, Exit 49 (River Street exit) off Route 495, Haverhill, Massachusetts. **Time:** 7:45 AM (Coffee and donuts along River Street). No size limit to trip but car-pooling is strongly encouraged. ACCESS TO ALL STOPS HAS BEEN GRANTED BY PERMISSION OF THE PROPERTY OWNERS; DO NOT ENTER IN THE FUTURE WITHOUT SEEKING PERMISSION FIRST!!

Directions for meeting along field trip route; depart from assembly point at 8:00 AM; proceed 1 mile south on 495 to exit 48 (Ward Hill Route 125) to STOP 1 (Haverhill Resource Recovery Facility; Ogden Martin Systems of Haverhill, Inc.). From 495 northbound take exit 48 and follow road to light at Riverview Street, U-turn and drive 0.6 miles north to facility; from 495 southbound take exit 48 and at top of exit circle proceed straight to stop sign and turn left; 0.1 miles to plant entrance. STOP 2 at approximately 9:00-9:15 AM in rear of student/staff parking lot Northern Essex Community College (NECC). From Stop 1 proceed 4.6 miles north on 495 to exit 52 (Haverhill Route 110), 0.2 miles on exit ramp and turn left (south) on Route 110; proceed 0.1 miles and turn left on unnamed street which leads to NECC (entrance sign on right side); proceed 0.3 miles to main entrance of NECC and assemble in rear of lot. First stop in Maine will be at Great Hill in Eliot off Route 101 west from Route 1 (Refer to Delorme's Maine Atlas and Gazetteer Map 1 or USGS Kittery, York Harbor, and Dover East 7.5-minute topo sheets).

Leaders of each field trip stops or significant contributor to stops are indicated by initials or asterisk next to initials.

### MILEAGE

Cum.	Int.	
miles	miles	

0.0		Westgate Shopping Center K-Mart, Exit 49 (River Street exit) off Route 495, Haverhill, Massachusetts.
-----	--	---

3.2	3.2	STOP 1. Leave plant, turn left onto access road and proceed to and lights.
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STOP 1 - HAVERHILL RESOURCE RECOVERY FACILITY, THE NECK, HAVERHILL, MASSACHUSETTS (BDS\*, TKW).

The large excavation is in a lowland area of thick till, surrounded by numerous drumlins with crests above 250 ft altitude (Figure 5; Haverhill, Massachusetts 7.5-minute quadrangle). The excavated hill is elongate but is irregular in shape and morphology; it is not a drumlin. The Canton fine sandy loam series was mapped on the original landform; Paxton very stony, or extremely stony, fine sandy loam was mapped on the drumlin west of Bare Meadow Brook (Figure 5), coincident with the area of mixed till shown in figure 5. There, drumlin till and thin mixed till are exposed in a large cut. Charlton and Canton very stony fine sandy loam series were mapped on Ward Hill and the other drumlins (Fuller and Hotz, 1981; Fuller and Francis, 1984).



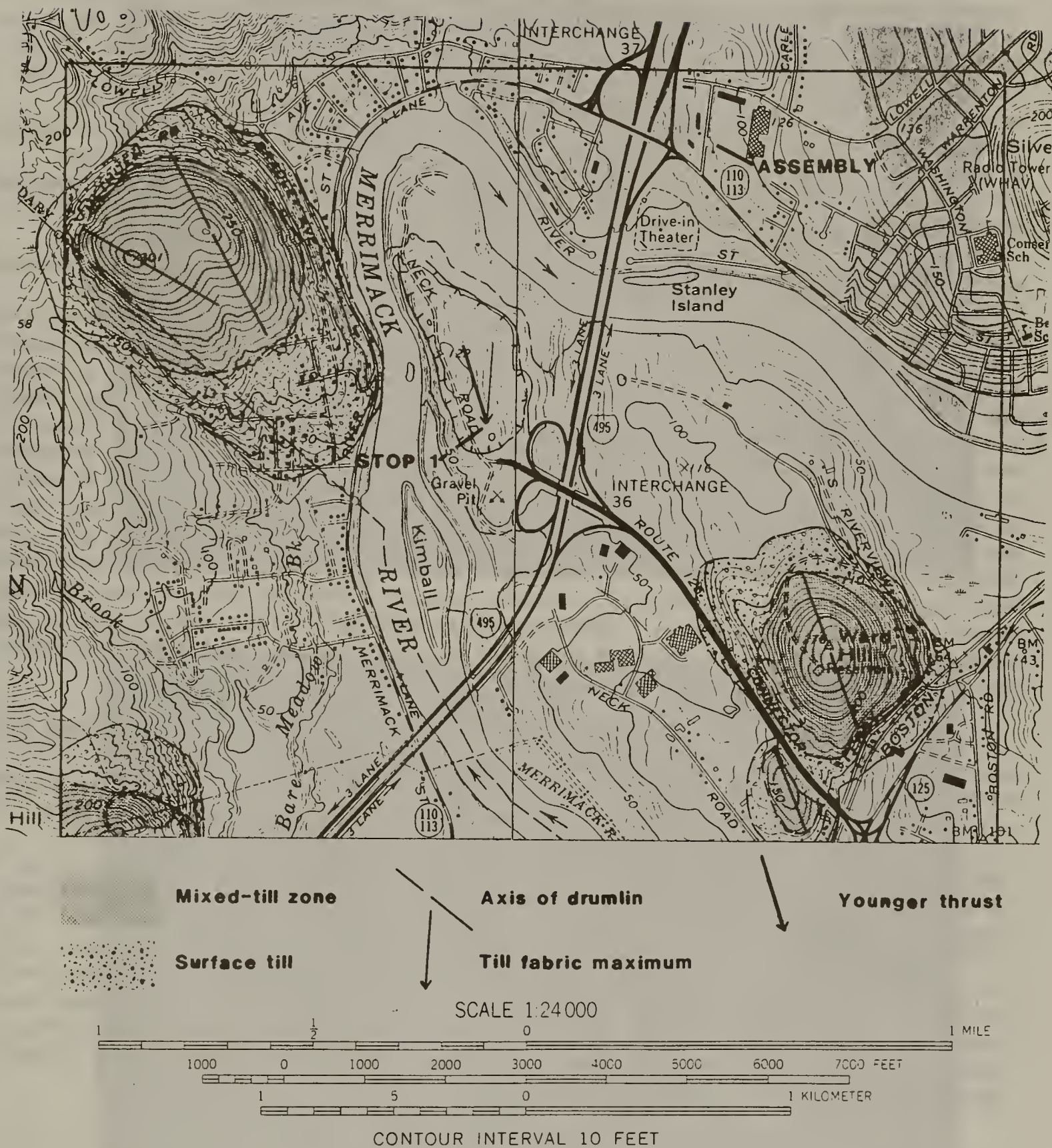


Figure 5. Map showing distribution of selected till units and location of Stop 1; base map Haverhill 7.5-minute quadrangle, Massachusetts, 1972.



The pit exposes 3-6 m of gray sandy till in bench cuts at the west end. The base of the lowest cut is about 40-50 ft below the top of the original landform. The till is light brownish gray to light gray (2.5Y 6/7-2), compact, thinly layered silty sand till with 15% gravel and boulders by volume. It is composed of gray to light brownish gray (2.5Y 6/0-2) massive, compact, nonplastic diamict material in irregular patches and lenses, <2 m long, which are associated with zones of gravel clasts (15-20% pebbles and cobbles). The bulk of the till comprises laterally extensive zones of alternating thin (<3 cm) layers of gray, massive, compact, silty-sand diamict material and thin layers of light gray to pale yellow (2.5Y 7/2-4) less compact sandy material. The massive material contains minute, and irregularly shaped, angular pebble-to-cobble-sized clasts of compact gray clayey silt (Figure 6a). These clasts are composed of indistinctly laminated and microlaminated gray clayey silt and white medium-to-coarse silt. Microlaminae are bounded by sharp contacts; laminae are folded in truncated open or isoclinal folds. Many of the silt clasts and gravel clasts are oriented with long axes dipping northerly -<20° (Figure 7a). Layering also dips northerly (Figure 7b). Sand and silty-sand layers are deflected around small pebbles and cobbles; individual layers cannot be traced much more than 20 cm. Sandy layers pinch out at or are truncated by the silty-sand layers. The till weathers and fractures along the layering. Sets of layers that dip as much as 50-60° are truncated by other sets that dip Generally 15-20°. Some steeply dipping resistant gray silty-sand diamict layers can be matched across the gently dipping planes of truncation, indicating thrust displacement of a few centimeters. Gravel clasts and boulders as much as 2.5 m long are subangular with subrounded ends or faces, and with angular, hackly terminations on some faces. Some faces of large boulders retain glacial polish and grooves. Till stones composed of fine-grained metasediments show some evidence of glacial shaping and many are striated.

The till is interpreted as a basal lodgement facies of the surface till, based on the degree of compaction, strongly preferred orientation of till stones, and preferred orientation of layering. The till-stone fabric records a southerly flow during deposition of the lower part of the exposed till. The till contains clasts of laminated silt that preserve primary laminae, indicating a source component of older stratified materials in the valley or materials deposited subglacially and reentrained during till sedimentation. Layers also contain minute clasts of the laminated silt. The layering of the till is very well developed here, and it is a regional characteristic. It is of probable subglaciogenesis, perhaps related to some primary segregation of material, which was enhanced by shear or thrust displacement along the sandy layers. Rotation of sets of layers took place during till sedimentation. The youngest, planar sets of thrusts show movement toward the southeast (Figure 7b), parallel to the axes of surrounding drumlins.

4.0    0.8    U-turn at lights.

4.5    0.5    Entrance to 495N.

9.8    5.3    Exit from 495 at Exit 52 (Haverhill Route 110) and turn left (south) on Route 110; proceed 0.1 miles and turn left on unnamed street which leads to NECC (entrance sign on right side); proceed to main entrance of NECC and assemble in rear of lot.



(a)



(b)

Figure 6. Photo details of surface till, Stop 1. a) Sheared clast of gray clayey silt in layered sandy till matrix; thin oxidized rim around clast, U.S. quarter for scale. b) Steeply dipping older thrust faults and layering, truncated by younger thrust; U.S. quarter for scale.



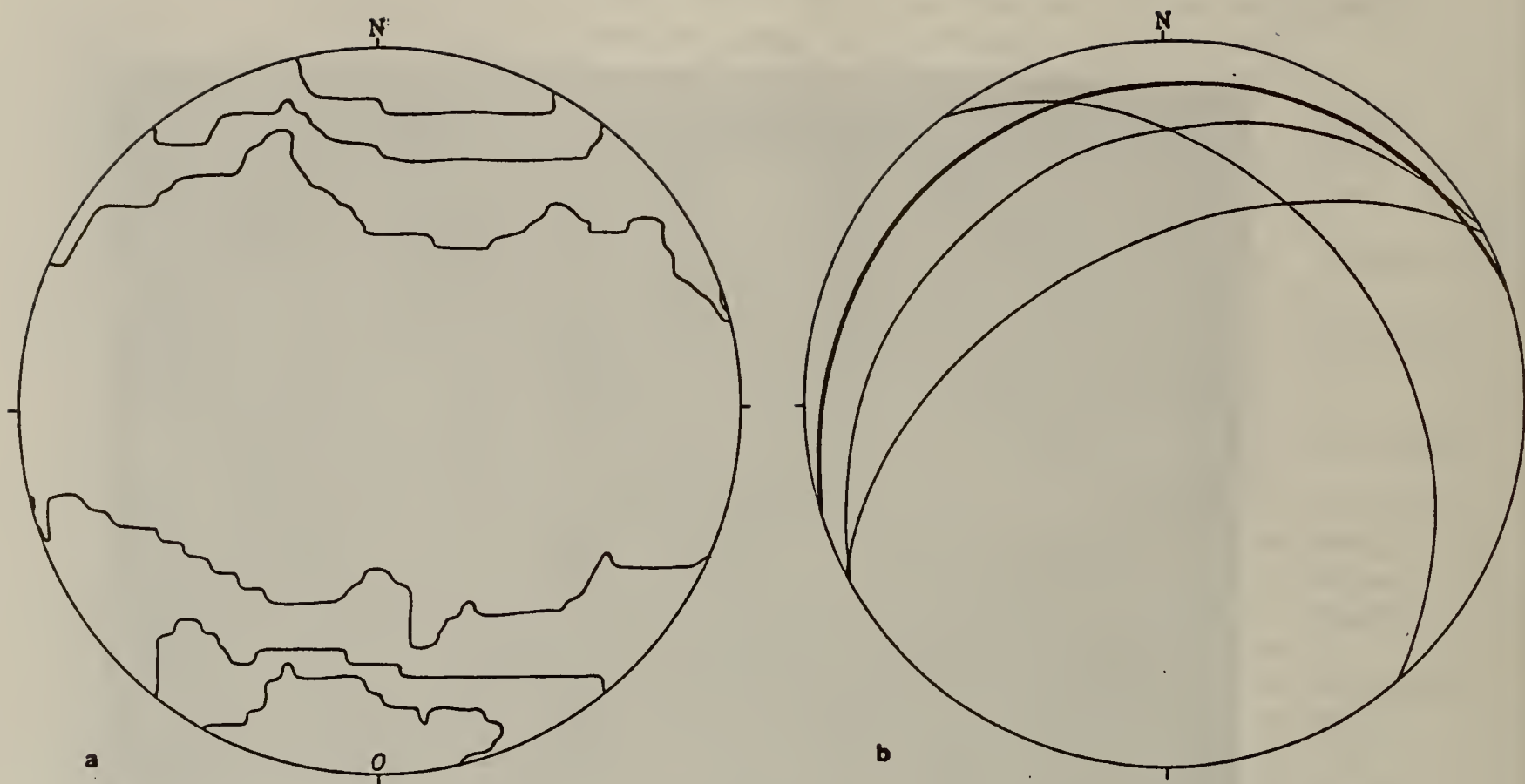


Figure 7. Kinematic elements of surface till at Stop 1; lower hemisphere projections. a) Fabric of 35 till-stone long axes, clasts exhumed from lower face of exposure; principal eigenvector trends  $186^{\circ}$ , plunges  $4^{\circ}$ ; contour interval 2.0 sigma, counting area 0.105. b) Older thrust faults and younger thrust fault (heavy line) from upper face of exposure; younger thrust indicates movement toward  $163^{\circ}$ .

- |      |     |  |
|------|-----|--|
| 10.7 | 0.9 | Leave NECC parking lot, turn left on Kenoza Street.  |
| 11.6 | 0.9 | Fork left on Center Street (Walnut Cemetery on right).   |
| 11.7 | 0.1 | T-junction, turn left.   |
| 11.8 | 0.1 | Turn right onto Millvale Road.   |
| 12.5 | 0.7 | Junction with East Broadway (bear right).  |
| 13.0 | 0.5 | <b>STOP 2.</b> Turn into entrance to pit on right (WATCH FOR TRUCKS!). Pull into pit and park. |

STOP 2 - MAIROFRIDES BROTHERS SAND AND GRAVEL PIT, EAST BROADWAY STREET,  
HAVERHILL, MASSACHUSETTS (BDS\*, TKW).

This pit exposes till in the sides and top of an irregularly shaped, double crested drumlin. Paxton extremely stony, or very stony, fine sandy loam soils were mapped in areas coincident with the mixed-till zone shown in figure 8 (Haverhill, Massachusetts 7.5-minute quadrangle) . The pit also exposes a thick section of glaciomarine deltaic beds, normal to paleocurrent flow, and younger glaciomarine silt and sand to the north (Figure 8). The fine sediments are unconformably overlain by crossbedded medium-coarse pebbly sand, of fluvial-glaciofluvial origin.

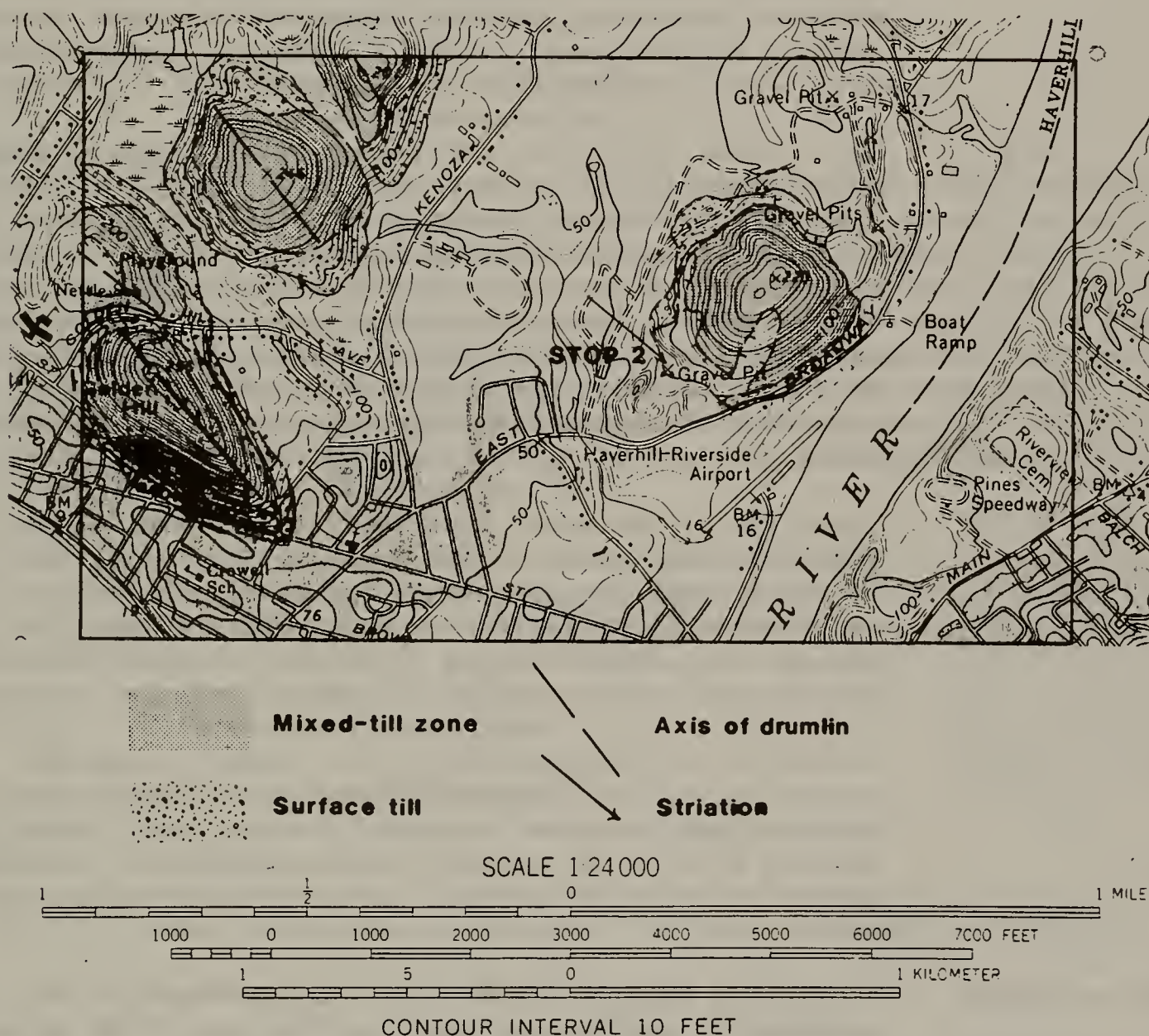


Figure 8. Map showing distribution of selected till units and location of Stop 2; base map Haverhill 7.5-minute quadrangle, Massachusetts, 1972.



The west-facing cut exposes an upper set of units that contain about 10% pebbles, cobbles, and few boulders. These units overlie an oxidized stone-poor till which grades downward into gray till. Bedrock crops out beneath the gray till on the floor of the pit. A generalized section is described, from top to bottom:

<u>Unit</u>	<u>Thickness</u>	<u>Description</u>
1	1-2 m	Pale yellow (2.5Y 7/4) and strong brown to reddish yellow to yellowish brown to brownish yellow (7.5-10YR 5-6/8) silty sand diamict material, noncompact to moderately compact, massive, containing pebbles, angular blocks and subrounded cobbles and boulders (10-25%); loose and easily excavated at north end of exposure; irregular base with 1 m relief.
2	2 m	Pale yellow (2.5Y 7/4) silty-sand till, compact, massive; unit is gradationally more stony from bottom to top (to about 10-15% but is less stony than overlying unit; pale yellow till matrix contains angular clasts of compact silty till; faces of these clasts are stained strong brown (7.5YR 5/8) with Fe-Mn coating; subangular gravel clasts include weathered brown metapelites; clasts are not coated with Fe-Mn stain; vesicles are common in parts of the matrix; base of unit is irregular, distinguished by color and stone content.
3	2-3 m	Olive (5Y 4-5/4) silty fine-sand till consisting of discontinuous layers of olive silty-sandy till, less than 2 cm thick, that contain subangular-to-blocky compact clasts of clayey-silty till; till clasts are less than 3 cm long and are coated with strong brown (7.5 YR 4/6) Fe-Mn stain; pebbles and cobbles are 2-5%; base of unit is locally sharp.
4	2-4 m	Olive (5Y 5/4) clayey-silty till; compact, massive, subhorizontal and subvertical joints well developed at top and more widely spaced downward through unit; joint faces stained with Fe-Mn coating; unit is mottled at base where it grades by color and loss of joint structure into underlying unit.
5	1-3 m	Gray till, compact, massive, tough, moderate to high plasticity, pebbles and small cobbles are 2-5%; stones composed of fine-grained rock types are subrounded and striated, but not well faceted.

## Stone Lithologic Compositions at STOP 2:

Dark gray to black, fine-grained  
metapelite (local bedrock).....28%  
Granite, granitic gneiss,  
pink, gray, and white.....26%  
Gray, fine-grained  
biotite schist.....15%  
Gray amphibolite, garnet  
schist, and gneiss.....12%  
Gray, fine-medium grained  
gneiss, hornblende(?),  
muscovite, biotite.....11%  
Gray, very fine-grained  
aplite (volcanic?),  
weathered to silt.....4%  
Quartz.....3%  
Brown, weathered biotite  
schist of metapelite.....1%

Unit 1 is interpreted to be chiefly of ablation origin, based on its loose, cobbly character, and degree of soil development. However, an early postglacial colluvial origin cannot be ruled out. Units 2 and 3 contain oxidized and fresh clasts of the underlying units within a sandy matrix that is similar in texture to the overlying unit and to the surface till at Stop 1. These units are stony varieties of the mixed-till zone; genetically they are local varieties of the surface till, the materials locally eroded and deposited on the drumlin skin by the late Wisconsinan ice sheet. Unit 4 is the pervasively oxidized and jointed weathering zone developed in the top of unit 5, the drumlin till of Illinoian age.

13.5	0.5	Leave pit, turn left onto East Broadway and retrace route back to Millvale Road.
14.3	0.8	Left onto Millvale Road.
15.1	0.8	Bear right at Middle Road.
16.2	1.1	Pass under 495.
16.6	0.4	Stop; T-junction with Route 110, turn left (WATCH TRAFFIC!).
17.6	1.0	Junction with 495N (2nd entrance) turn left and proceed on 495N.
23.7	6.1	Drumlin on right; oxidation profile visible in excavation.
27.2	3.5	Junction with I-95N.
28.6	1.4	New Hampshire - Massachusetts border.
34.3	5.7	Hampton/Exeter toll (\$0.75); proceed on 95N.



- 41.9 7.6 Follow 95N to Portsmouth to all Maine points, and proceed across Piscataqua River over Portsmouth/Kittery Bridge.
- 44.5 2.6 Maine - New Hampshire border.
- 45.8 1.3 Exit off 95 at Exit 3; stay in right lane for Route 236.
- 45.9 0.1 Bear right to Route 236.
- 46.2 0.3 Turn right onto Route 236 at yield sign (flashing red lights).
- 50.6 4.4 Turn right at Depot Road (school on right).
- 51.2 0.6 Stop; turn right.
- 51.3 0.1 Immediate left into Great Hill Fill and Gravel Inc.
- 51.5 0.2 **STOP 3.** Park to right of vertical face; after discussion drive 0.3 miles up to top of drumlin.

STOP 3 - GREAT HILL DRUMLIN, ELIOT, MAINE (DWC\*, JMC\*, TKW)

Stop 3 is at a large excavation into Great Hill, off of Goodwin Road, in Eliot, Maine (Figure 9) The topographic maps that cover the site include the Dover East 7.5-minute and Dover 15-minute quadrangles. The surficial geology of the Maine portion of the Dover East quadrangle was mapped by Smith (1988). Note that this is not the Kennebunk Great Hill drumlin from which the 13,800 date was obtained. Something has to be done regarding the proliferation of Great Hills in Maine.

Great Hill is one of several drumlin forms present in the southeast corner of the Dover East quadrangle and in adjoining quadrangles to the south and east. Smith (1988) has mapped DeGeer moraines on the flanks of several of the drumlins in the area, while the lows between drumlins are underlain by Presumpscot Formation silts and clays. Most of the drumlins were eroded and modified by wave activity, and sand and gravel deposits formed during marine regression are found at elevations below 200 feet, the local limit of marine submergence.

Caldwell and others (1979) reported that deformed sand is found in the northwest ends of several drumlins in this area. They reported that the stratified drift in the drumlins is capped by unweathered till and was assigned a Late Wisconsinan age. Furthermore, because of similarities between the stratified drift in the drumlin core and that found in nearby end moraines, they suggested the stratified drift in the drumlins may represent older moraines, or stratified drift deposited on the flanks of the drumlin, which was later incorporated into the drumlins by readvance. In this instance, Caldwell and others (1979) suggested the Kennebunk readvance may have been responsible for the stratified drift in the drumlins. Smith (1981), however, has shown that the Kennebunk readvance was most likely due to minor ice-marginal oscillation and not a significant climatic event. While this does not preclude incorporation of stratified drift into the drumlins by ice advance, the presence of an oxidized horizon noted during the summer of 1989 on the top and flanks of the drumlin in the till over the stratified drift suggests its incorporation may be due to an ice advance older than the Late Wisconsinan.



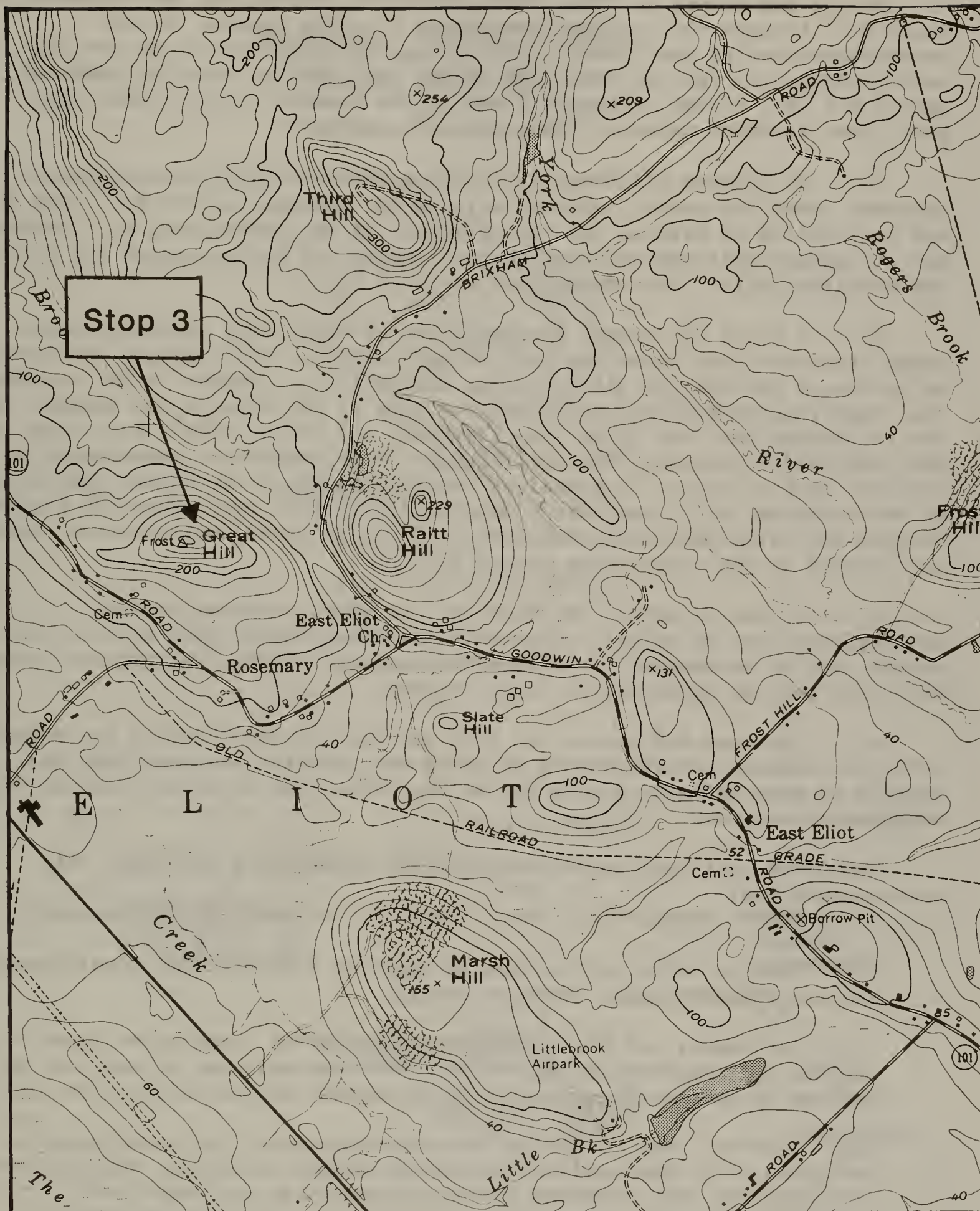


Figure 9. Map showing location of Stop 3 (Great Hill drumlin); base map Dover East 7.5-minute quadrangle, Maine-New Hampshire, 1973.



During the summer of 1989, three different excavations were present at Great Hill (Figure 10), an inactive borrow pit (#1) located at the southeastern end of the drumlin, with a near-vertical face 30 meters high, an active borrow pit (#2) at the crest of the hill, and an inactive sand and gravel pit (#3) on the northeast flank of the drumlin. Within these three pits, four distinct deposit lithologies are exposed:

1) An unknown thickness (3 meters exposed) of glaciotectonically deformed sand and gravel, exposed only in the highwall of Pit #1. This unit was not studied in detail, due to the danger of the overhanging till exposure, but it appears that the sand and gravel in part may be the source for much of the overlying till. This deposit is overlain by

2) A thick (25 meter) stratified to massive, very compact, coarse sandy diamicton that forms the bulk of the deposit. This unit is interpreted in part as a deformation till, grading upward into a lodgment till. A spine of the underlying stratified drift from the core of the drumlin was present in the pit during the 1984 NEIGC trip (Mayewski and Birch, 1984), but it has now been completely removed. The till appears to be completely unweathered in the vertical face of pit #1, but shallow excavations along the access road to pit #2 show oxidized till, and over 5 meters of oxidized, pale olive (5Y 6/4) very compact till with prominent manganese staining on joint and tillstone surfaces are exposed in the active face of pit #2.

3) A thin (1 meter) ablation till is poorly exposed in the southeastern face of pit #1. This unit resembles the surface till in the region, and it is assumed to be Late Wisconsinan in age. The contact between this till and the drumlin till is not exposed.

4) The sand and gravel pit (#3) was badly slumped during the summer of 1989. It exposes several meters of sands and gravels, reworked from the drumlin by wave activity during marine recession. No contacts with the other units were observed.

Questions that need to be discussed and answered by the field trip participants are:

1) Is the drumlin till the "lower till" of southern New England?

2) What criteria can be used to prove or disprove this identification, and how diagnostic are these criteria?

3) For years, pit #1, which displays only the unweathered drumlin till was the only exposure at this site. What criteria can be used in the absence of the pit #2 exposures to determine the identity of the till?

4) Finally, what is the age and significance of the underlying sand and gravel? If the sand and gravel is deltaic in origin, what could have impounded the body of water into which the delta was deposited?

Stop 3, NEIGC 1989  
Great Hill, Eliot

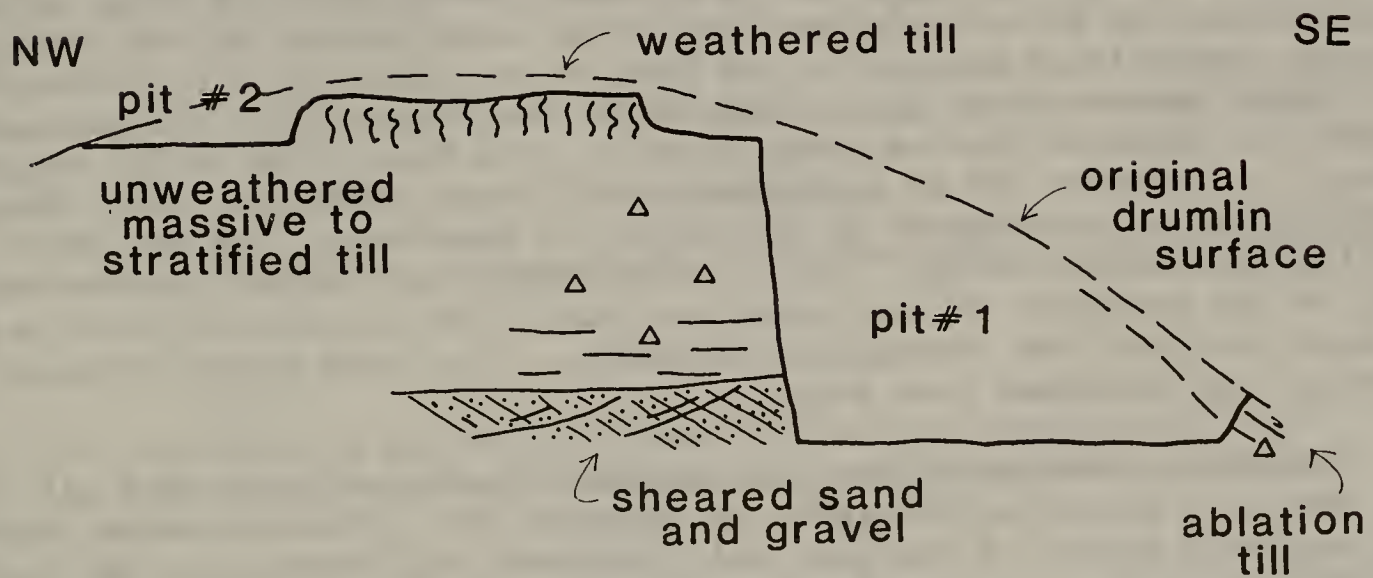
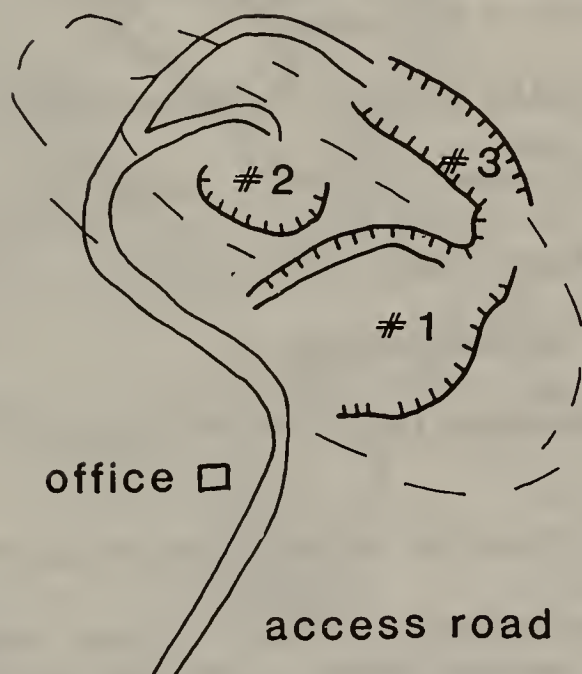


Figure 10. Pit locations and generalized geologic relations at Stop 3.



- 52.3 0.5 Leave top of drumlin, turn left (east) at exit gate onto Route 101 (Goodwin Road).
- 57.3 5.0 Stop; T-junction with Route 1, turn right (south) at lights.
- 57.5 0.2 Bear right to 95N and proceed north.
- 63.0 5.5 York Toll Booth (Exit 1 Maine Turnpike), take ticket, proceed north on 95/Maine Turnpike.
- 74.7 11.7 Exit off turnpike at Exit 2 (Wells/Sanford \$0.40).
- 75.2 0.5 Junction with Route 109, turn (west) right to Sanford.
- 82.7 7.5 Left onto Airport Road.
- 83.1 0.4 Turn left at end of fence.
- 83.3 0.2 **STOP 4.** Turn right into pit and park.

STOP 4 - SANFORD AIRPORT (LION HILL) DRUMLIN, SANFORD, MAINE (JMC\*, TKW)

Stop 4 is at a borrow pit on Airport Road, immediately northwest of Sanford Municipal Airport, Sanford Maine. The pit is located approximately 0.4 miles (0.6 km.) south of State Routes 4A and 109, on the east slope of Lion Hill (Figure 11). The topographic maps that cover this site include the Alfred 7.5-minute and Kennebunk 15-minute quadrangle maps. The surficial geology of the area has been mapped by Smith (1977).

Lion Hill is a small drumlin, approximately 0.55 km. long and 30 meters high, with a long axis trend of 136 degrees, consistent with other drumlin orientations and striation directions in the neighboring Sanford quadrangle (Clinch, unpublished mapping). The drumlin is located at the southeastern end of a long, bedrock cored upland separating the Mousam River and extensive sand and gravel deposits located along Route 4. The base of the hill, and the floor of the borrow pit is approximately 250 feet above sea level, about the limit of marine submergence in the region as determined from the elevation of glaciomarine deltas nearby. One of these deltas is located north of Route 109, to the northeast of the borrow pit site. The airport is sited on a layer of sand overlying the Presumpscot Formation. This sand marks the upvalley limit of the Kennebunk Sand Plain.

Two till lithologies have been observed in the walls of this pit (Figure 12, sketch of pit relationships), an ablation till 3+ meters thick, exposed in the southeast corner of the pit, and a lodgment till exposed on the high, graded, western wall of the pit. The contact between these two till lithologies has not been observed, but if the crude stratification within the ablation till is projected upward, it is reasonable to assume that the ablation till overlies the lodgment till.

The ablation till exposures display a wide variety of thin to thick, interbedded units, including sandy diamictons and sheared silts and sands, interpreted as minor debris flows, and gravel lenses, silt and sand beds, and



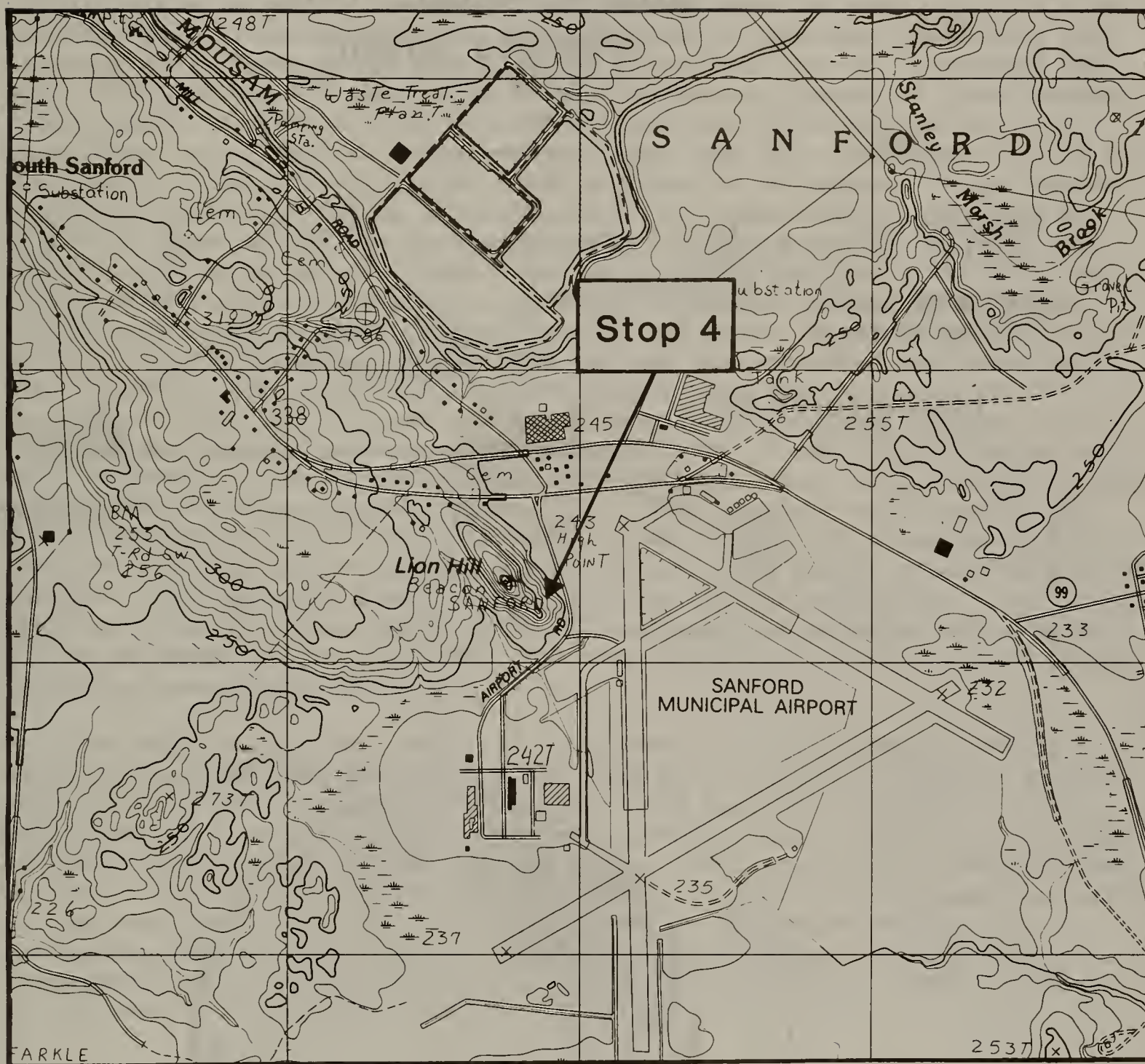


Figure 11. Map showing location of Stop 4 (Sanford Airport/Lion Hill drumlin); base map Alfred provisional 7.5-minute quadrangle, Maine, 1983.

channel sands, indicative of fluvial reworking during deposition. Small (2-3 cm) inclusions of the underlying weathered lodgment till are included within the sandy layers of the till. This unit is pale yellow (5Y 7/3-7/4), largely unweathered and unoxidized, with only a weak, postglacial soil profile present. The age of this unit is inferred to be Late Wisconsinan, and the till resembles typical exposures of the surface ablation till found in much of southern Maine and the crystalline highlands of southern New England, the "upper till".



# Stop 4, NEIGC, 1989 Sanford Airport

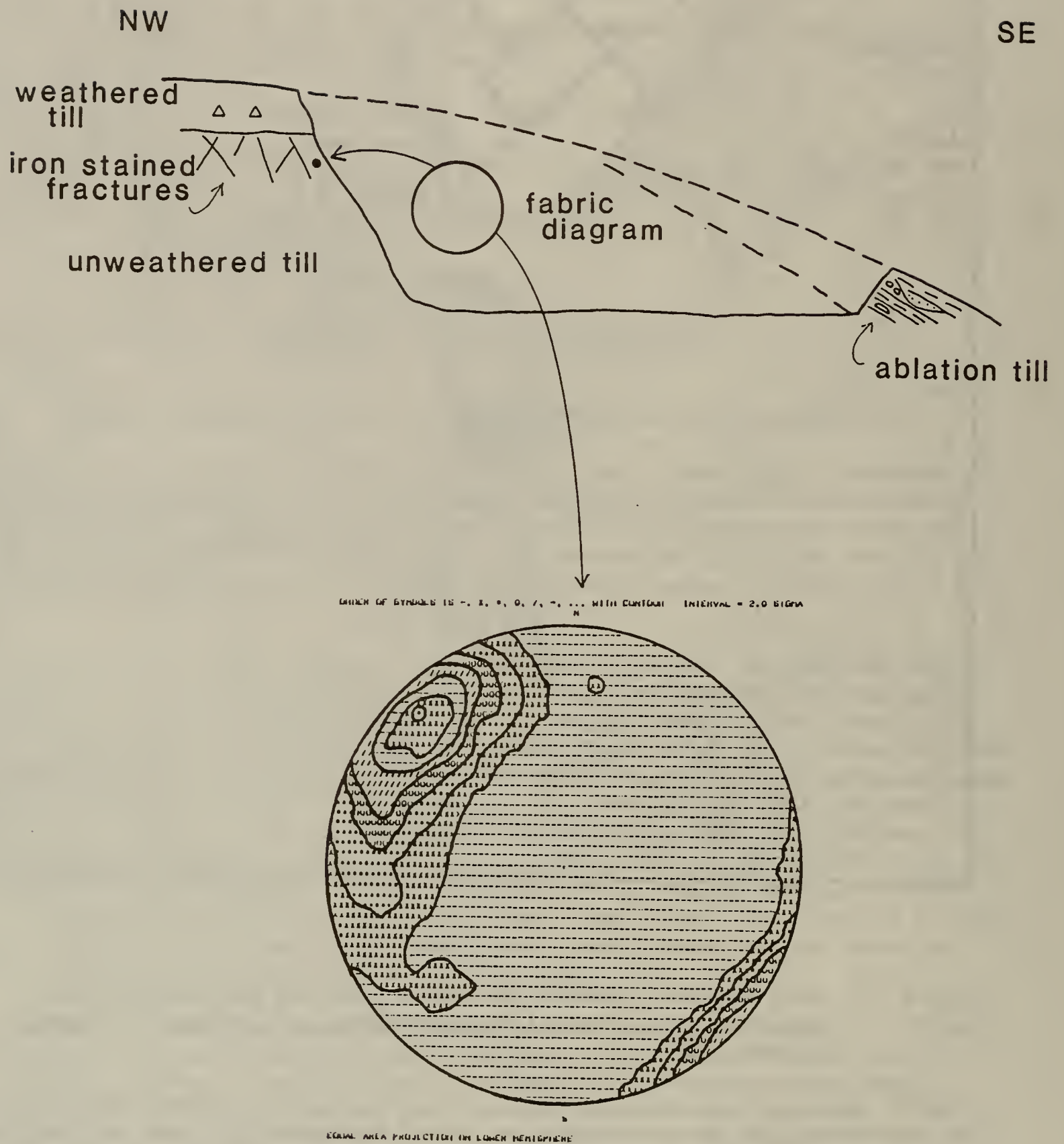


Figure 12. Generalized geologic relations and petrofabric from core of drumlin at Stop 4 (fabric data  $n = 50$ ,  $S_1 = 0.72$ ,  $V_1 = 303$ , 18).

The western face of the pit, at which the lodgment till is exposed was extensively graded and reworked prior to the summer 1989 field season, and only a limited amount of exposure could be excavated by hand. At this site, a compact, silty to sandy matrixed diamicton with a prominent weathering zone was observed. Three grain size analyses taken from this till averaged 60% sand, 23% silt and 17% clay, with little variation between the three samples and no apparent difference between the texture of the weathered and unweathered till. Fabric measurements, taken from the unoxidized till at this site gave a trend of 303 degrees, with a plunge of 18 degrees (significance value, or measure of fabric strength, is 0.72, where random is 0 and 1 is uniform. These values are consistent with a lodgment origin, and with the long axis of the drumlin. A one meter thick, olive to pale olive (5Y 5/4-4/4 moist, 5Y 6/3 dry) oxidized zone, with discontinuous, dark reddish brown (2.5YR 3/6), sandier zones is developed in the olive gray to gray (5Y 5/2-4/1 moist, 5Y 6/2 dry) till. The contact between the weathered and unweathered till is sharp, but rusty, iron stained joints have been observed in the unweathered till for at least one meter beneath the contact, and the till is oxidized for about one cm on either side of these joints. Following a heavy rain and several days of dry weather, the stained joints were still wet, while the surrounding unweathered till was dry. This suggests that some of the joint staining may still be forming today, as a result of groundwater flow. When the site was better exposed, prior to 1989, the upper weathering zone was reported to be several meters thick, the iron stained joints were also observed beneath the weathered till, and staining had occurred along these joints outward from them into the till matrix for several centimeters in the otherwise unweathered till.

The thickness of the weathered zone observed during previous years suggests that the lodgment till in the core of the Lion Hill drumlin may be correlated to the "lower till" of southern New England. Samples of both weathered and unweathered till have been taken for grain-size and clay mineral analysis, although all of the results are not available at the time this was written. However, without the benefit of the earlier observations made before the pit was graded what criteria are present to determine whether the weathering zone is the result of interglacial weathering (the "two-till" interpretation) or a post-glacial, groundwater permeability effect of no age significance?

ALTERNATE STOP (No mileage logged) - PEPIN AND SONS PIT, SANFORD, MAINE (JMC)

Stop # 5 is at a shallow excavation at the Pepin and Sons sand and gravel processing site (marked as a sand pit on the topographic map) in Sanford, Maine. The pit is located approximately 0.4 miles (0.6 km) northwest of Goodall Memorial Hospital. Topographic map coverage of the site includes the Sanford 7.5-minute and Berwick 15-minute quadrangles. The glacial geology of the Sanford quadrangle is currently being mapped by Clinch (in prep.).

The pit is cut into the northwest end of a 1.5 km long drumlin, with a long axis trend of 136 degrees. Dissected kame terraces and deltas are present southwest and west of the drumlin, while a thin veneer (0.5-1.5 m.) of very bouldery, sandy ablation till overlies bedrock to the northeast.



Two till lithologies are exposed in the northern wall of the pit; here about 2 meters of ablation till overlies an unknown thickness of lodgment till. The lodgment till is a very compact, coarse sandy to pebbly diamicton with well-developed fissility and jointing. The till appears to be somewhat oxidized, with a light yellowish brown to pale olive color (2.5-5Y 6/4), with prominent iron and manganese staining on joint and tillstone surfaces. No unoxidized till has been observed at this site. The contact between the lodgment till and the overlying ablation till is sharp.

The ablation till is a crudely laminated, yellow (2.5Y 7/8) sandy diamicton, estimated to be at least 2 meters thick, although the upper portion of the unit has been disturbed and graded. Small clasts and slabs of weathered lodgment till are present in the lower 20 cm of the ablation till, and some of the sand and silt layers appear to have been sheared. A wedge of sandy ablation till material that dipped in the glacier flow direction was observed in the lodgment till during the summer of 1989.

This exposure can be interpreted as a "two-till" locality, with a Late Wisconsinan "upper till" overlying an Illinoisan "lower till" exposed in the core of a drumlin. However, none of the diagnostic criteria that would prove a two glaciation origin for the exposure are present. The lodgment till is oxidized, but not to a great extent and the ablation till is also oxidized. Inclusions of lodgment till and sheared sands in the ablation till, and the filled wedge in the lodgment till are present, but a mixed zone of weathered and unweathered lodgment till was not observed. Any interpretation of this exposure as a "two-till" site rests solely on non-diagnostic criteria, such as geomorphic relationships, till facies, texture, compactness and extensive manganese staining. Your opinions on how much (or little) weight should be given to these criteria at this and other sites would be welcome.

- |      |     |   |
|------|-----|---|
| 83.5 | 0.2 | Leave pit, turn left and proceed to stop sign. Turn right onto Airport Road.  |
| 84.0 | 0.5 | Stop sign, bear right (east) back onto Route 109.   |
| 84.6 | 0.6 | Junction with Route 99, turn left (WATCH TRAFFIC!). Proceed on Route 99 over Kennebunk Plains.  |
| 90.6 | 6.0 | Turn left at Maine Turnpike sign.   |
| 90.7 | 0.1 | Stop; turn left at stop sign.   |
| 90.9 | 0.2 | Cross Mousam River.   |
| 91.3 | 0.4 | Stop; turn right at stop sign (to Maine Turnpike/95N).  |
| 92.4 | 1.1 | Left turn at turnpike entrance. Proceed through toll booth to Turnpike/95N but turn left at Service Area for lunchstop.                         |
| 92.9 | 0.5 | <b>LUNCHSTOP.</b> Left turn to Service Area (Burger King); discount mileage. Please return to vehicles no later than 45 minutes after stopping. |



- 93.2 0.5      Leave lunchstop and enter turnpike northbound. Passengers may sleep off lunch.
- 141.2 48.0    Exit from turnpike at Exit 12 (Auburn \$1.60). Proceed through toll booth.
- 141.5 0.3      Stop; turn left (north) at stop sign onto Route 4/202/100.
- 144.9 3.4      **STOP 5.** Turn left across highway (WATCH TRAFFIC!). Park in back of lot.

**STOP 5 - NESS OIL COMPANY, AUBURN, MAINE (MJR)**

The exposure at this stop is located off the southbound lane of Washington Street (Route 4-100) in Auburn, shown on Figure 13 (Lewiston 7.5-minute quadrangle). The surficial geology of the Lewiston 15-minute quadrangle which includes the area has been mapped by Smith and Thompson (1980).

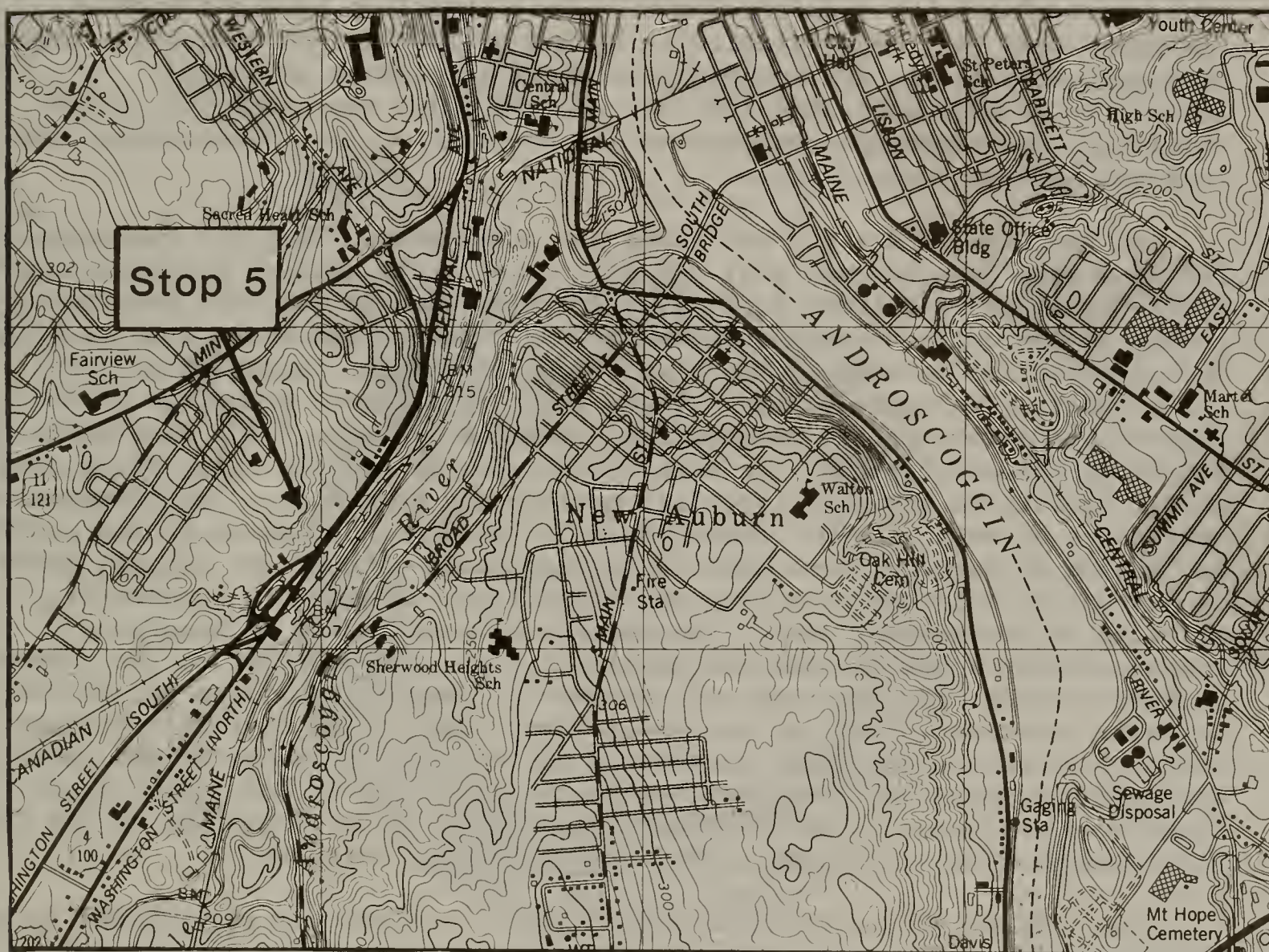


Figure 13. Map showing location of Stop 5; base map Lewiston 7.5-minute quadrangle, Maine, 1979.



At one time there was approximately 50 to 60 feet (15-20 m) of near vertical exposure at this site, however continued slumping has forced the owners of the property to grade back the upper sections of the pit and build a retaining wall along part of the foot of the slope. Most of the section, however, is visible in steps up the face.

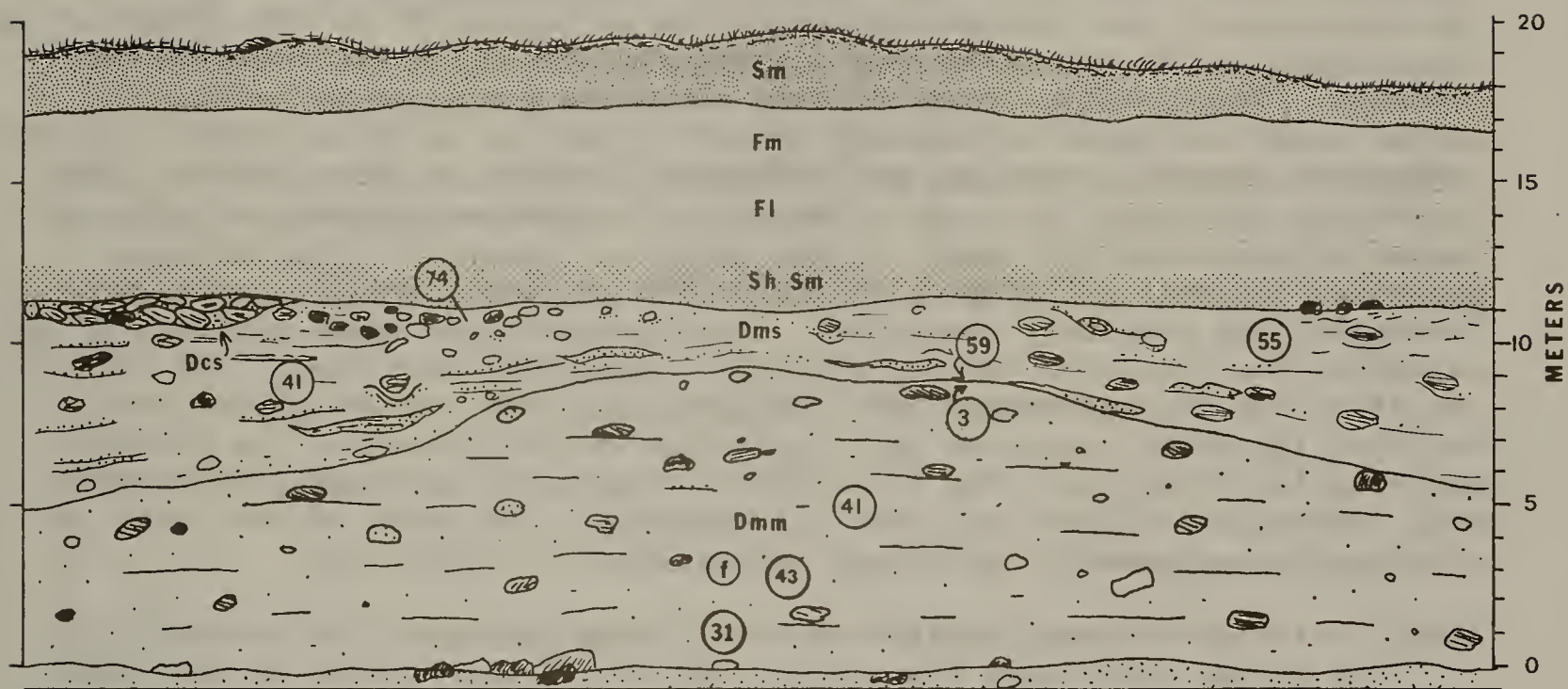
The exposure at this stop is situated on the south facing slope of a modified drumlinoid hill approximately 100 feet high that is part of a bedrock topographic high south and west of the city of Auburn and northwest of the Little Androscoggin River. The long axis of the hill is oriented approximately 155 degrees; local striae average close to 170 degrees. The base of the site is at an elevation of approximately 200 feet; the crest of the hill is approximately 300 feet. The local marine limit has been estimated at 336 feet (Thompson and Borns, 1985a) in the Mt. Auburn Delta located 3 miles north of the site, however, the late Wisconsinian marine submergence may have extended as much as 5 miles inland (north) of the site.

Figure 14 is a stratigraphic log through approximately 20 m of exposure in several levels of the face. The stratigraphic sequence in this exposure includes two diamictons (Dmm, Dms) near the base and middle of the exposure (which are of primary interest for this trip) overlain by a sequence of laminated sand (Sh, Sm), rhythmically bedded silty clay and sand (Fl), massive fine-grained deposits (Fm), and thick-bedded to massive sand (Sm). Collectively, the upper sand and fine-grained units represent the marine inundation and regression associated with retreat of late Wisconsinian ice from the coastal lowland zone.

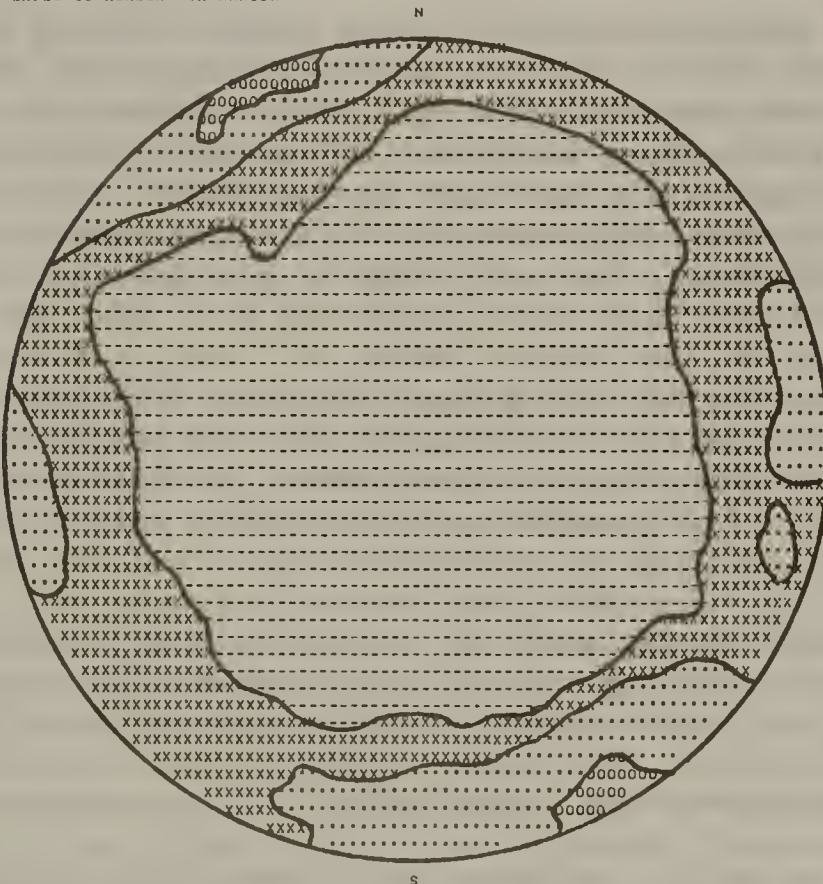
At the base of the pit and along the paved access road for approximately 200 m, a very compact dark gray (5Y 4/1; damp) silty massive matrix-supported diamicton is exposed (Dmm, Fig. 14). The diamicton is thickest near the center of the outcrop and thins towards the flanks. Clasts in this material are striated and polished. The clast compositions are dominated by local lithologies. The most abundant clasts are calc-silicate granofels of the Taylor Pond member of the Sangerville Formation (Hussey, 1983) that underlies the hilly terrain north of the site. Other clasts include rusty weathering pelitic schist and grainite, granitic gneiss, and pegmatite. The fine-grained matrix has a range in sand content from 31% to 43%. Oxidation of the matrix is limited to vertical fractures in the uppermost zone of the deposit where it is in contact with the overlying sandy diamicton unit and is apparently affected by groundwater flow. The lower diamicton is cut by numerous fractures that resemble sheeting joints that are sub-parallel to the present outcrop surface, dipping east and southeast. Several fractures dip in the opposite direction, that is into the outcrop face. A clast fabric (Figure 14, n=36), measured on the lower section of the deposit is weakly bimodal with the strongest orientation lying parallel to the direction of the local striae (regional ice-flow). A weaker clast orientation is perpendicular to ice flow with clasts lying within the shallow east-dipping features.

The compact gray diamicton is overlain by a diamicton of variable texture and composition that thins over the crest of the underlying deposit and thickens towards the flanks of the outcrop. The contact with the underlying gray unit is distinct and abrupt and is represented by a color change from gray to olive and a change in grain size and structure. In general, the

# Stop 5 , Lewiston NEIGC 1989



36 POINTS       $n_{\text{est}} = 1$   
COUNTING AREA = .100  
EXPECTED NUMBER = 3.60  
STANDARD DEVIATION (SIGMA) = 1.80 = (EXPECTED NUMBER) / 2.0  
ORDER OF SYMBOLS IS -, X, \*, O, /, -, ... WITH CONTOUR INTERVAL = 2.0 SIGMA =  
EXPECTED NUMBER \* IX RANDOM



EQUAL AREA PROJECTION ON LOWER HEMISPHERE

Figure 14. Schematic diagram showing geologic relations (above) and petrofabric (below) from lower diamicton (letter f in circle) at Stop 5. Fabric data  $n = 36$ ,  $S_1 = 0.51$ ,  $V_1 = 325$ ,  $0.2$ . Numbers in circles denote % sand in matrix of diamicton at location shown on text; lithologic symbols are given in text.



material in this unit is a compact olive (5Y 5/3) stratified, matrix-supported diamicton (Dms, Fig. 1). The matrix of the unit is texturally variable across the outcrop with sand contents ranging from as low as 40% in the center of the unit to 77% in the uppermost zone of the unit underlying the glaciomarine sequence. The diamicton contains thin contorted and disrupted medium to coarse sandy stringers and several layers of sand up to 40 cm thick. The sand stringers contain laminae and are commonly oxidized and iron stained. Some clasts are surrounded by rings of medium to coarse sand. Isolated pods and lenses of sand also are common in the deposit. Clasts in this unit are dominated by local lithologies. Some of the pelitic rocks (Sangerville Formation) are intensely weathered. Locally within the upper sections of the exposure, the diamicton is very stony. In uppermost sections of the outcrop the diamicton is clast-supported (Dcs, Fig. 1). Many of the clasts are striated flatirons, oriented with flat sides on the planes of the prominent east-dipping foliation. The foliation in some cases is seen as outlined by thin discontinuous fine sand and silt layering. The upper contact with the glaciomarine sequence is an abrupt truncation.

After visiting the previous stops several questions should be apparent:

- (1) Is the lower gray compact unit at Stop 5 correlative to the "lower till" of southern New England?
- (2) Is the upper olive unit at Stop 5 correlative to the "upper till" of southern New England? Alternatively;
- (3) Are the two diamictons products of one glaciation, representing a lodgement facies (gray till) and either subglacial meltout till or a debris flow diamicton deposited in a proximal proglacial setting?

Although the sequence exposed in the Auburn site is similar in some aspects to two till localities in southern New England, we believe that there are several important distinctions that favor a model for two lithofacies within one glaciation. Here we focus the question on the lower diamicton unit. As we will see in the sections at New Sharon, grain size of the diamictons (in this case lodgement till) is a function of parent material that the glacier advances over, not antiquity of the deposit. In the case of the Auburn site, we believe that the glacier may have advanced over proglacial materials deposited in the lowland basin in front of the advancing ice sheet. Whether these materials are (1) primary erosion products from subglacial erosion of the local pelitic bedrock, or (2) advance of ice through a pre-late Wisconsinan marine unit or (3) a lake dammed in the valley of the pre-glacial Androscoggin Valley, fine-grained materials were incorporated by the ice sheet.

A second aspect is the lack of weathering (oxidized horizon) in the upper zones of the lower diamicton. As we have seen earlier today, an olive to olive brown zone is visible on top of many lower till sections in southern New England, but is absent at this site. Clay mineral analysis (cf. Newton, 1978) may address this problem more accurately (work in progress).

145.2 0.3      Leave Stop 5 and turn left (north) onto highway. The next 3 miles will be through downtown Lewiston, so PAY ATTENTION AND GARDEZ LA DROIT!!!

- 145.9 0.7      Junction with Routes 121/11. Stay on right side of main road (Route 4N).
- 146.3 0.4      Turn right at lights (Route 4 continues north, but you turn right onto Route 202/11/100). Follow main road.
- 147.4 1.1      Junction of 126 with 202/11/100 (bear to left); Kora Shrine Temple on right.
- 148.2 0.8      Lights (junction of 202/11/100 with Russell Street); continue north on main road.
- 149.2 1.0      Lewiston Raceway on right. Continue northward.
- 166.3 17.1     Bear right on Route 41/133.
- 168.0 1.7      Bear right on Route 41 (to Readfield and Mt. Vernon).
- 168.1 0.1      Stay right on Route 41 (very quick so watch it!)
- 172.8 4.7      Stop; junction of Routes 41 and 17, turn right on Route 17.
- 174.4 1.6      Turn right onto dirt road just before Readfield Depot.
- 174.5 0.1      **STOP 6.** Pit on right; turn into it and park.

**STOP 6 -MACE PIT, READFIELD, MAINE (WBT)**

This inactive pit is located in a small hummock that reaches an elevation of 97 m (Figure 15; Readfield 7.5-minute quad). It is just below the upper marine limit, and glacial marine silt (Presumpscot Formation) occurs nearby at similar elevations (90 - 94 m). The surficial geology of the Augusta 15-minute quadrangle which includes Stop 6 has been mapped by Thompson (1977). The pit face has exposed up to about 4 m of section, much of which is now concealed by slumping. Digging in this face reveals the following three till units:

Unit 1) The lowest unit is a very compact, silty, moderately stony till of probable lodgement origin. The stones are subangular to subrounded, and some are faceted and striated. Iron/manganese staining occurs on stone surfaces, but is only sparsely developed in the slightly fissile matrix of the till. Sand laminae are present, but are not as abundant as in unit 2. The till is oxidized, with a color intermediate between 5Y 4/2 (olive gray) and 2.5Y 4/2 (dark grayish brown). Unit 1 is correlated to the lower (and older) of the two principal tills examined on this trip, and thus is equivalent to the Nash Stream till of Koteff and Pessl (1985).

Unit 2) The middle unit is best exposed at the northeast end of the pit, where it is up to 1.6 m thick. It is a silty-sandy, moderately stony till, with an oxidized olive-colored matrix (5Y 5/3). This unit is loose and easily excavated due to its fissility and abundant sand laminae, which have promoted physical weathering and extreme development of fine blocky structure. The contact with unit 1 ranges from sharp to gradational over a few cm, and is



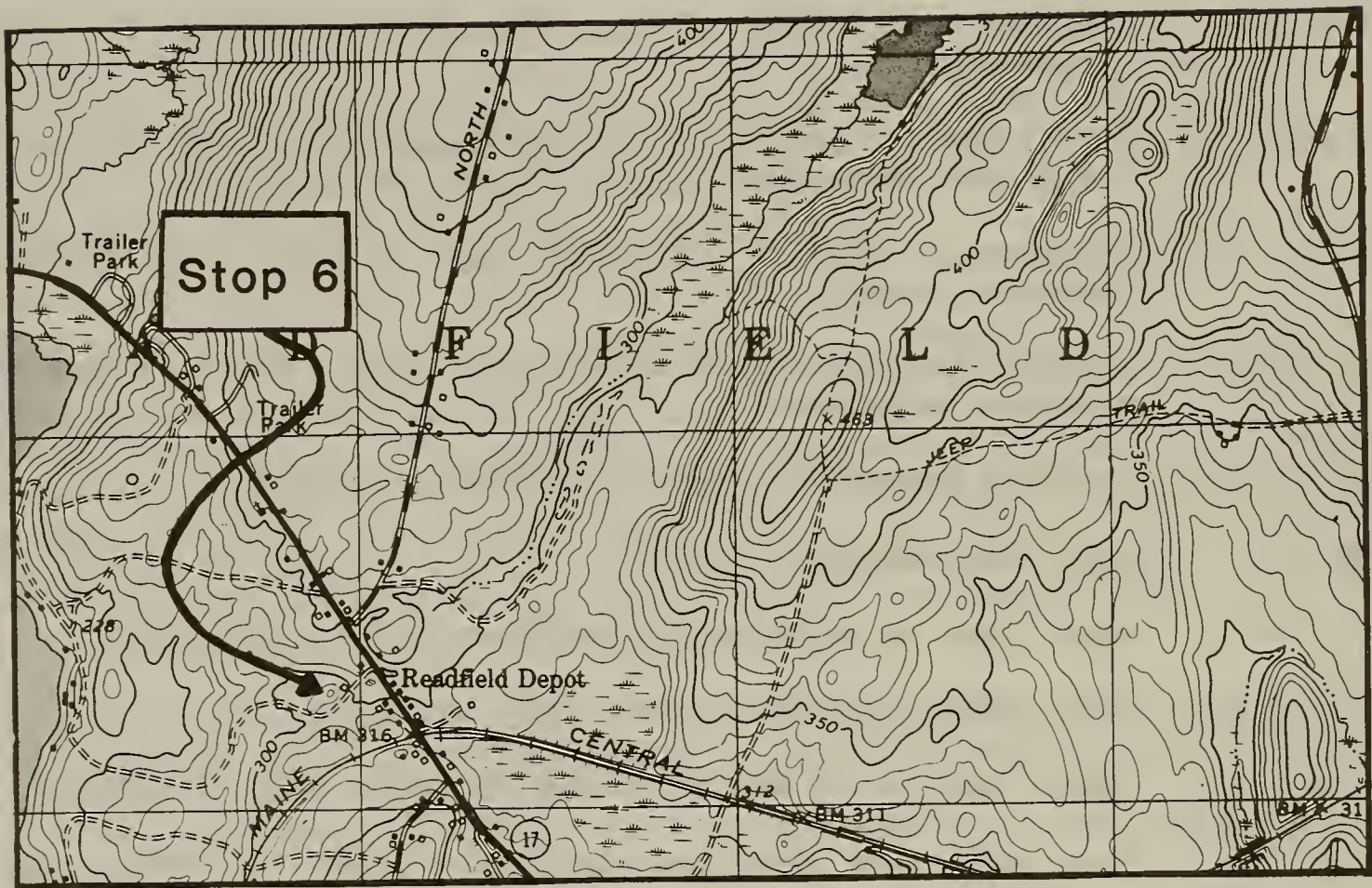


Figure 15. Map showing location of Stop 6; base map Readfield 7.5-minute quadrangle, Maine, 1980.

defined by the change in compactness and abundance of sand laminae. Two possible interpretations are suggested for this unit. It may simply be a sandy, fissile zone in till of the same age as unit 1, or it may be a basal mixed zone of the overlying Late Wisconsinan till (unit 3). In the latter case, the unit consists largely of sheared inclusions of lower till which have been glacitectonically mixed with sandy sediment from the basal part of the Late Wisconsinan till.

Unit 3) The uppermost unit consists of 0 - 1.7 m of variably oxidized, loose, sandy to gravelly till. It contains many stones up to 1 m in diameter, which are generally quite angular. A few stones are faceted and striated. Although this unit resembles some poorly sorted gravels that are derived from marine erosion of tills, it contains numerous clasts of silty till from unit 1 or unit 2. These clasts would not have survived if the deposit had been subjected to wave attack and extensive reworking. There are some imbricated stones in one part of the unit, but they may have been oriented by debris flow from glacial ice. Unit 3 is interpreted as an ablation facies of the Late Wisconsinan surface till that occurs throughout the region. It is the equivalent of the Stratford Mountain till of Koteff and Pessl (1985).

174.6 0.1      Leave pit and return to Route 17; turn left and retrace route back to 41/17 junction in Readfield.



- 176.3 1.7 Proceed straight through lights in Readfield (junction 41/17) to Kents Hill.
- 178.3 2.0 Bear right on Route 41 to Mt. Vernon.
- 185.5 7.2 Turn left onto Route 41 to Mt. Vernon.
- 192.6 7.1 Kennebec/Franklin County line.
- 193.0 0.4 Turn right onto Route 134 (dirt road).
- 196.8 3.8 Stop; junction with Route 2/27; turn right.
- 197.0 0.2 **STOP 7.** Pull over to right side of road and exit vehicles; we will cross the highway and walk through field on left to exposure on the Sandy River. LOOK BOTH WAYS AND CROSS HIGHWAY CAREFULLY; WATCH OUT FOR VEHICLES!!!!

**STOP 7. - SANDY RIVER EXPOSURE, NEW SHARON, MAINE (TKW)**

Stop 7 is located on Figure 16 (New Sharon and Mercer 7.5-minute quadrangles), and was referred to by Caldwell (1959) as locality D. The surficial geology of the Farmington 15-minute quadrangle which includes Stop 7 was mapped by Caldwell (1986).

The section exposed here is located on the southern bank of the Sandy River at New Sharon, downstream and on the opposite bank from locality C of Caldwell (1959), the site of the New Sharon Beds (Figure 16). The Sandy River flows northeastward here and the valley is flanked by highlands to the south which reach elevations up to 360 m asl.

Figure 17 is a schematic representation of the stratigraphy exposed here. Beneath 1 - 2 m of Holocene stream alluvium, about 4 m of fine-grained layered deposits comprised of interbedded fine sand, silt, clay, and diamicton. At the western end of the section, directly opposite the site of the New Sharon Beds, approximately 1.3 m of olive gray (5Y 5/2), compact silty sandy diamicton with striated clasts unconformably overlies the fine-grained deposits, which are strongly deformed. Also here between the fine-grained deposits and the sandy diamicton is a 1 m thick medium-grained sand, light yellowish brown (10YR 6/4) in color, which also is strongly deformed.

The layered sediments overlie a 4 m sequence of layered, gray (5Y 5/1) compact diamictons, which vary in thickness from 0.5 to 1 m thick, and in places are separated by thinly layered fine-grained deposits, usually less than a few cm thick comprised silt and clay layers and thin diamicton laminae. Clasts in the layered diamictons are subrounded to subangular and many are striated.

Under the layered diamictons is a 2 to 3 m thick, compact gray (5Y 5/1) to olive gray (5Y 5/2) clayey silt, with striated, subangular to subrounded pebbles and cobbles. Carbonate concretion nodules (some striated) are interspersed throughout the unit, and found in a "mixed zone" between the layered diamictons and the clayey silt. The unit was described by Caldwell (1959) as a "boulder clay", an appropriate characterization although it is



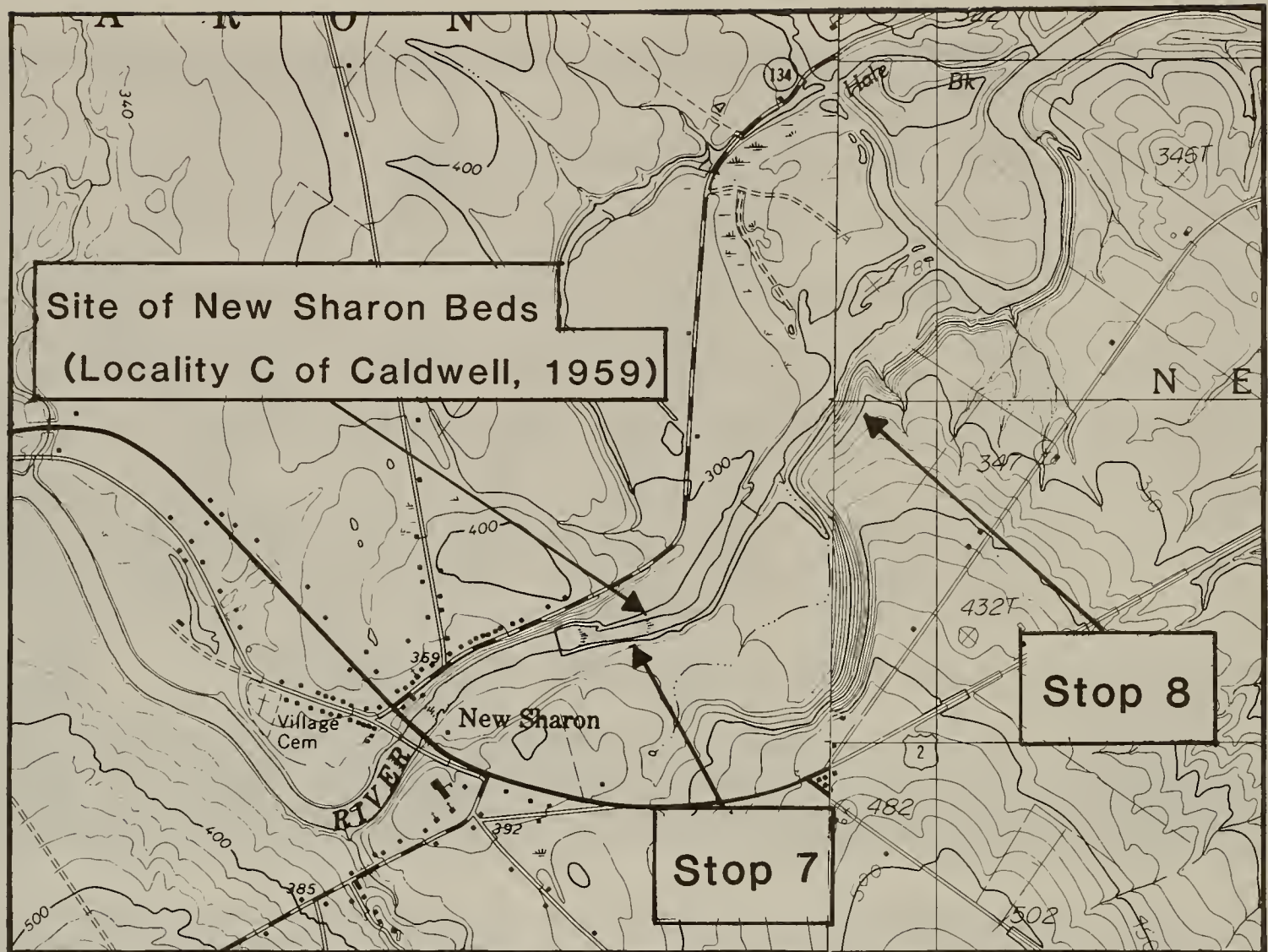


Figure 16. Map showing location of Stop 7 and 8, and location of the New Sharon Beds; base maps New Sharon (1968) and Mercer provisional (1982) 7.5-minute quadrangles, Maine.

more silty than clayey. The silt appears massive in its upper part, however, thin beds of dark gray (5Y 4/1) diamicton which pinch out laterally are found within the lower part of the unit, and can be seen as thicker lenses in the unit below river level. Downstream the silt becomes more stony and eventually grades into a crudely stratified diamicton which contains striated clasts. The base of the section is covered by slump and the river.

The carbonate concretions in the "boulder clay" have finite radiocarbon age-dates of about 32,500 and 40,000 yrBP, and an infinite date of >48,000 yrBP (Richard Pardi, pers. commun., 1989). Pardi (1989) has suggested that some concretions can be shown to be derived from organic matter deposited contemporaneously with the deposits they are found within, and that dates on these concretions may be applicable to the direct date of the enclosing sediment.

The "boulder clay" was tentatively correlated by Caldwell (1959) with the till he described as found beneath the New Sharon Beds, and the units above the "boulder clay" were correlated with the surface till in the region. The till described by Caldwell (1959, 1960) beneath the New Sharon Beds has not been encountered in recent excavation (Weddle, 1986), or seen in the river exposures since the lead author began work at New Sharon.

# Schematic Stratigraphic Section, Stop 7, New Sharon, NEIGC 1989

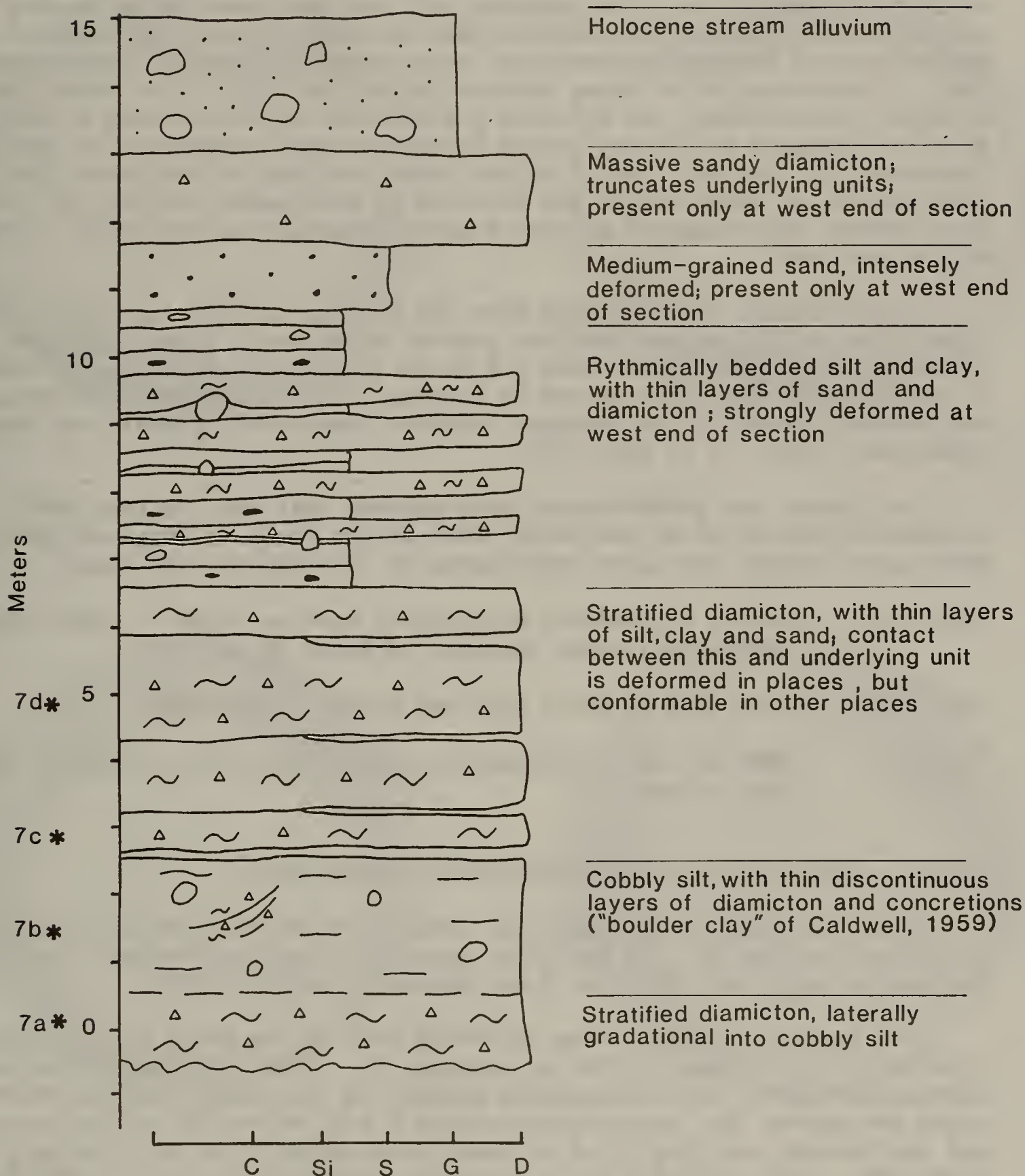


Figure 17. Schematic diagram of stratigraphy at Stop 7 (C = clay, Si = silt, S = sand, G = gravel, D = diamicton; 7a\* = petrofabric location).



The facies association and deformation at this section is complex. In general, the sediments reflect deposition by proximal ice into a proglacial lake, interrupted by ice-marginal fluctuations when lodgement or other basal processes deposited till, and the sediments were glacitectonically deformed by ice shove. The lower part of the section (0 - 10.5 m, Fig. 17) is comprised of fine-grained deposits of interbedded diamicton, silt, and clay. Graded beds, lateral continuity of fine-grained deposits, dropstones, and flow folds are present in these units indicative of the subaqueous origin of the deposits. Thicker diamictons are found between the fine-grained units (2.5 - 6.5 m, Fig. 17), and petrofabrics measured in some of the diamictons are indicated on Figure 18. The fabrics, all from the lower units, have a preferred NE-SW or N-S orientation, near perpendicular to the trend of striations and streamlined features in the region (Caldwell, 1986; Weddle, 1987). The strengths of these fabrics range from 0.57 - 0.78, some are bimodally distributed, and all have a preferred northward clast plunge. The fabrics also have relatively strong contour cluster suggestive of basal origin (especially 7a and 7d), but it is not clear how much of the stratified diamicton is of basal origin and which is of subaqueous debris flow origin. It is likely that examples of both types of deposits can be found at this part of the section.

Glacitectonic data measured from the section is shown on figure 19. These data can be grouped into two general sets; early planar features (bedding and thrusts) which have a N to NE dip and are found in the lower part of the section, and later NE and NW dipping planar features (fractures, folds, and thrusts) found in fine-grained deposits immediately beneath the uppermost diamicton (11.5 - 13 m, Fig. 17).

The fabric and glacitectonic data suggest that the ice flow was originally controlled by the NE-SW trend of the river valley, and that flow of this early sublobe was later overwhelmed by the regional ice flow.

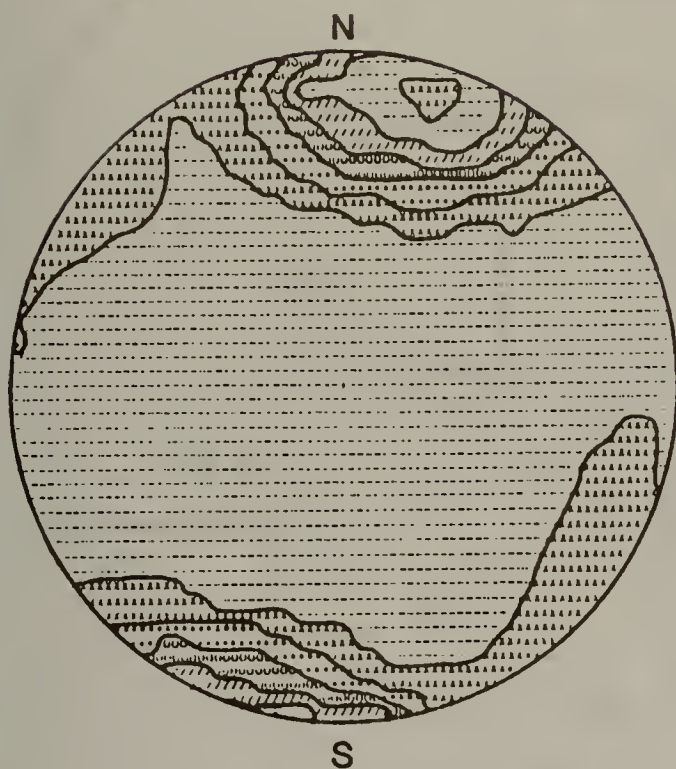
- |           |  |
|-----------|--|
| 197.5 0.5 | Return to vehicles and proceed east on Route 2. Bear left immediately after junction of Route 27 with 2. |
| 198.1 0.6 | Turn left at metal gate and proceed along road.  |
| 198.6 0.5 | <b>STOP 8.</b> Park vehicles and walk along road to exposure along the Sandy River.                      |

#### STOP 8 - SANDY RIVER SECTION, NEW SHARON, MAINE (TKW)

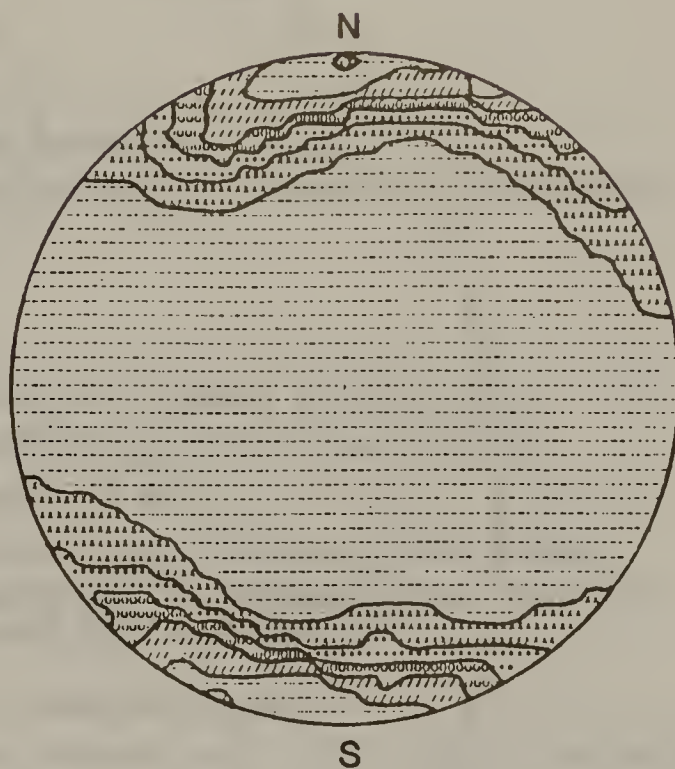
Stop 8 is located on Figure 16. This site was referred to by Caldwell (1959) as locality E. The surficial geology of the Norridgewock 15-minute quadrangle which includes Stop 8 was mapped by Weddle (1987).

The section is found along the south bank of the Sandy River and was referred to by Caldwell (1959) as locality E. Figures 20 and 21 are schematic representations of the stratigraphy exposed at this site. Section 8A occurs along the access road where approximately 1 m of olive (5Y 4/3) glaci-marine mud conformably overlies 0.5 m of light olive brown (2.5Y 5/6) compact sandy diamicton. This diamicton has a prominent subhorizontal platy fissility and contains at its base fragments and smeared beds of the underlying deposits.

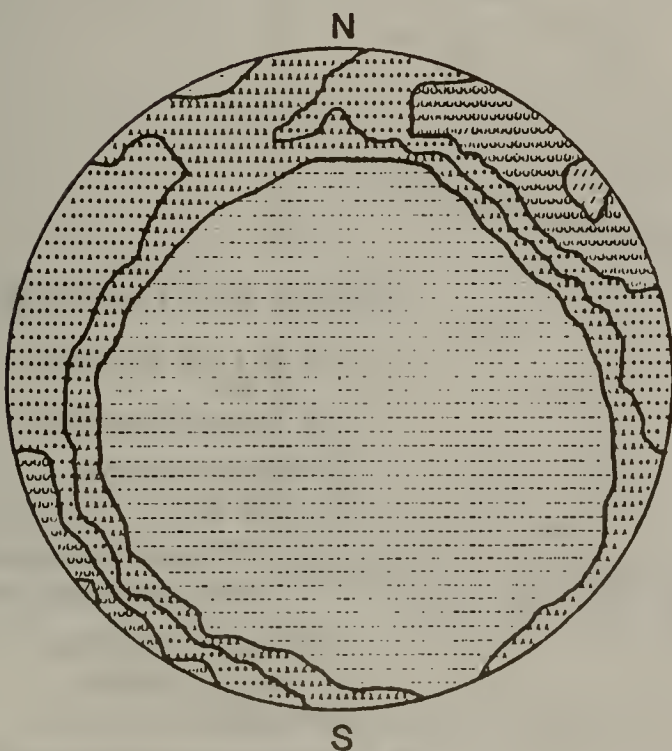
# Petrofabrics, Stop 7, New Sharon NEIGC, 1989



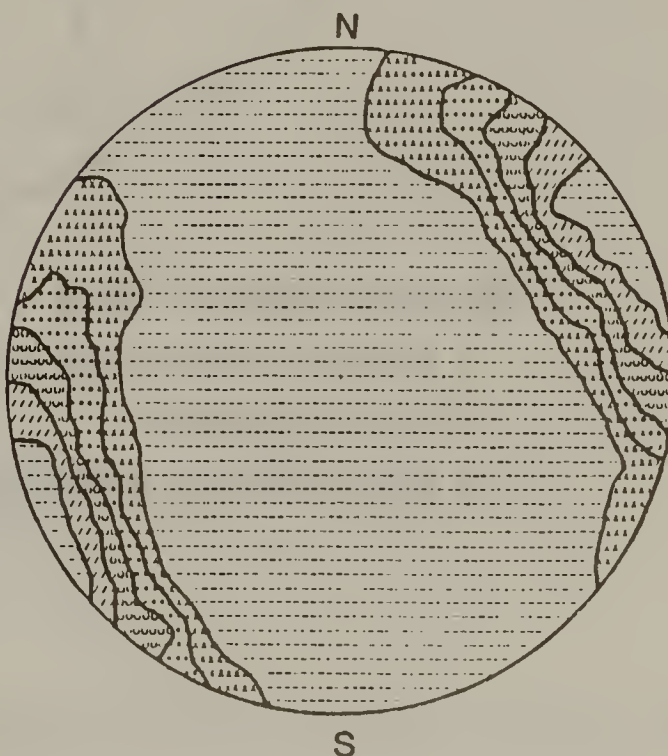
7a\*



7b\*



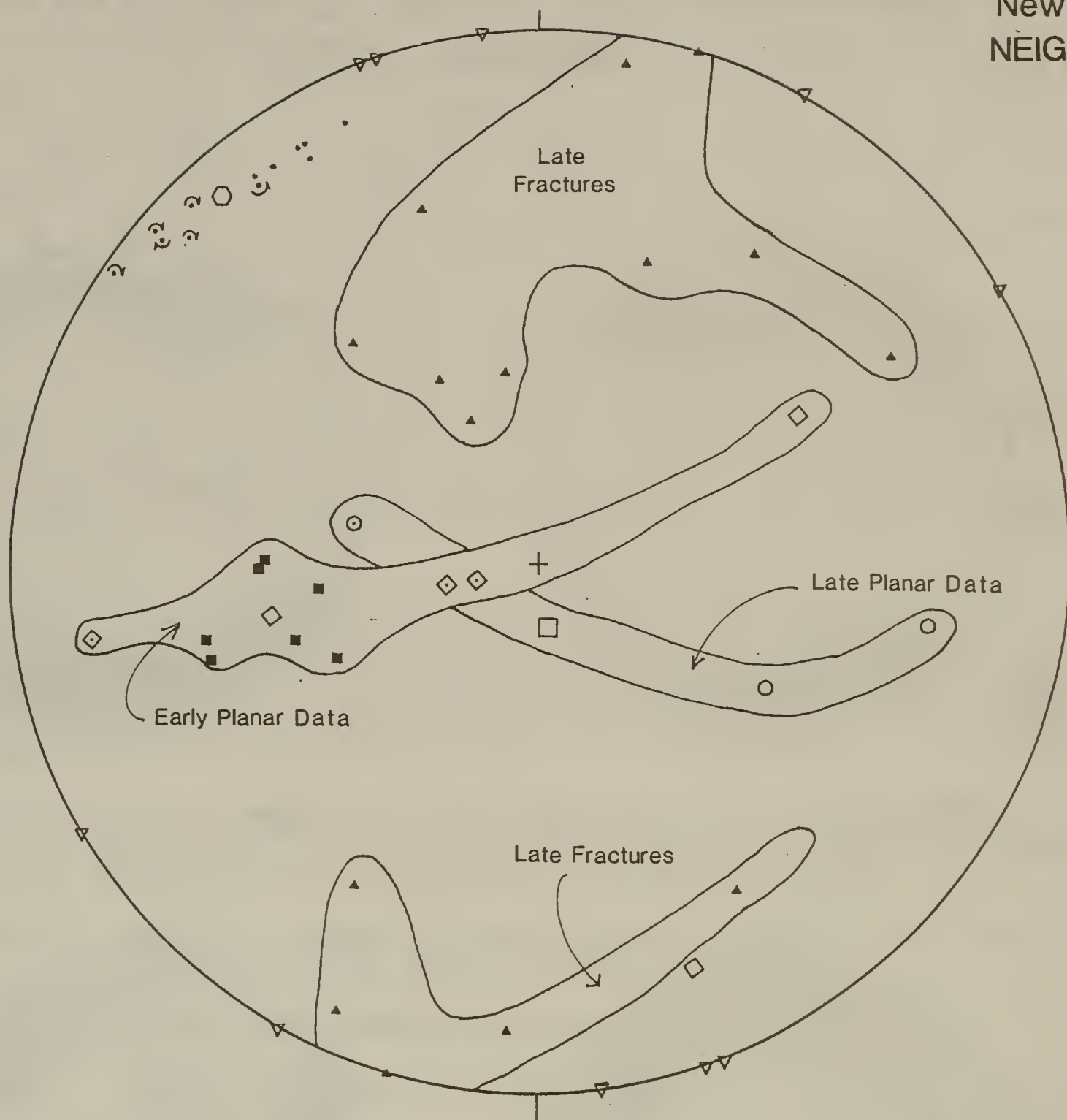
7c\*



7d\*

Figure 18. Petrofabrics from units at Stop 7; refer to Fig. 17 for locations (Fabric data 7a to 7d respectively,  $n = 34, 62, 50, 33$ ;  $S_1 = 0.75, 0.74, 0.57, 0.78$ ,  $V_1 = 11, 10; 186, 0.2; 55, 5; 65, 2$ ; contour interval = 2 sigma).





#### Slip-Line Analysis Data

- ↻ Fold Axis (rotation sense noted)
- Slip Line (N41W9)
- Pole to Slip Plane (N89E10NW)

#### Planar and Linear Data

- ▲ Pole to Late Fractures
- Late Planar Features
  - Pole to thrusts
  - ⊙ Fold AP
- ▽ Trend of Linear Features Observed in River (fractures ?)
- Early Planar Features
  - ◇ Pole to Thrusts
  - ◊ Fold AP
- Bedding (parallel to early thrusts)

Figure 19. Glacitectonic data from units at Stop 7; two general fields are present, 1) early NE-dipping planar features from lowest units (below 8 m on Fig. 17), and 2) later NE to NW-dipping planar features from higher units (above 8 m on Fig. 17).

# Schematic Stratigraphic Section, Stop 8A, New Sharon, NEIGC 1989

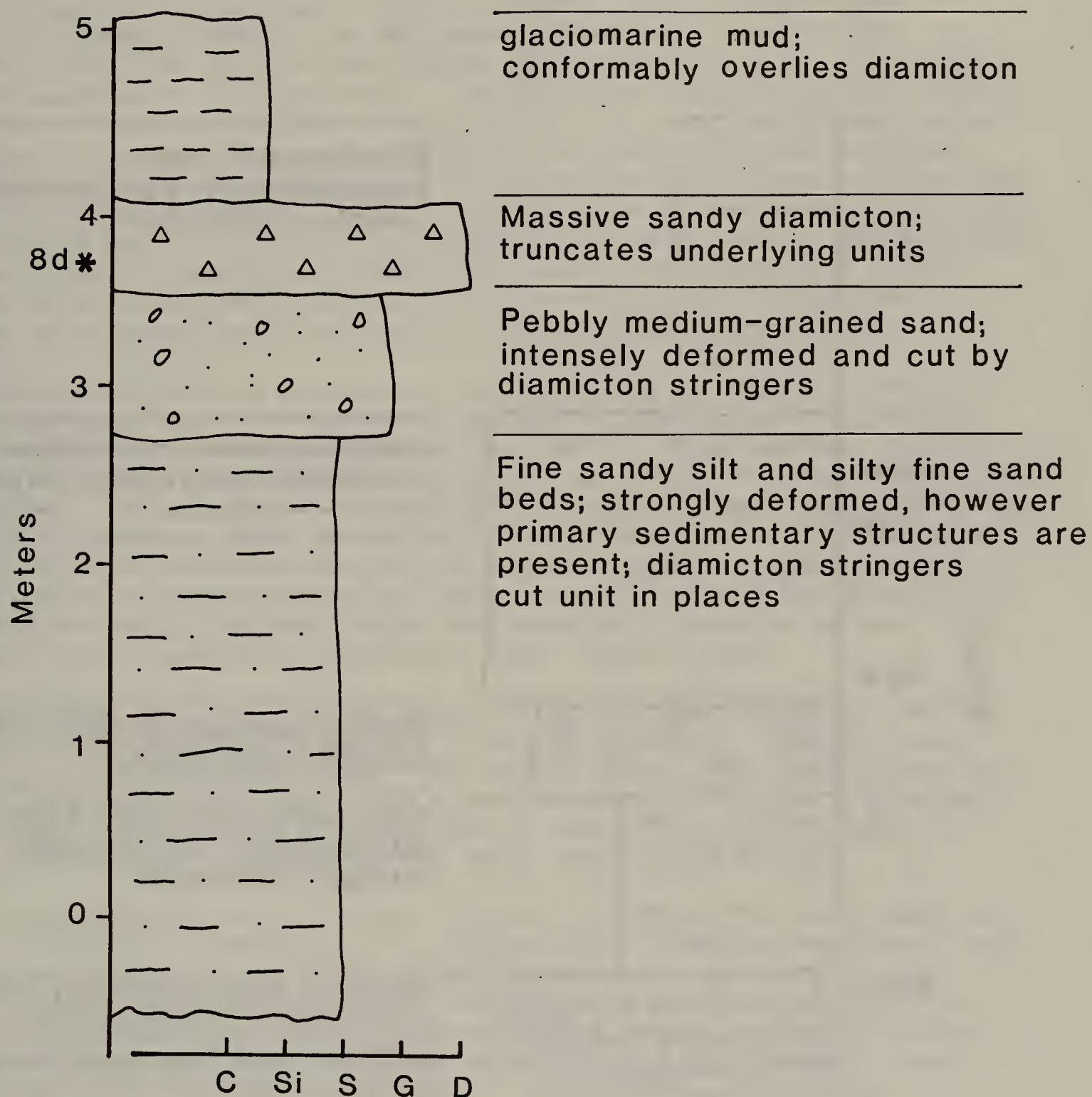


Figure 20. Schematic diagram of stratigraphy at Stop 8A (symbols same as Fig. 17).



# Schematic Stratigraphic Section Stop 8B, New Sharon, NEIGC 1989

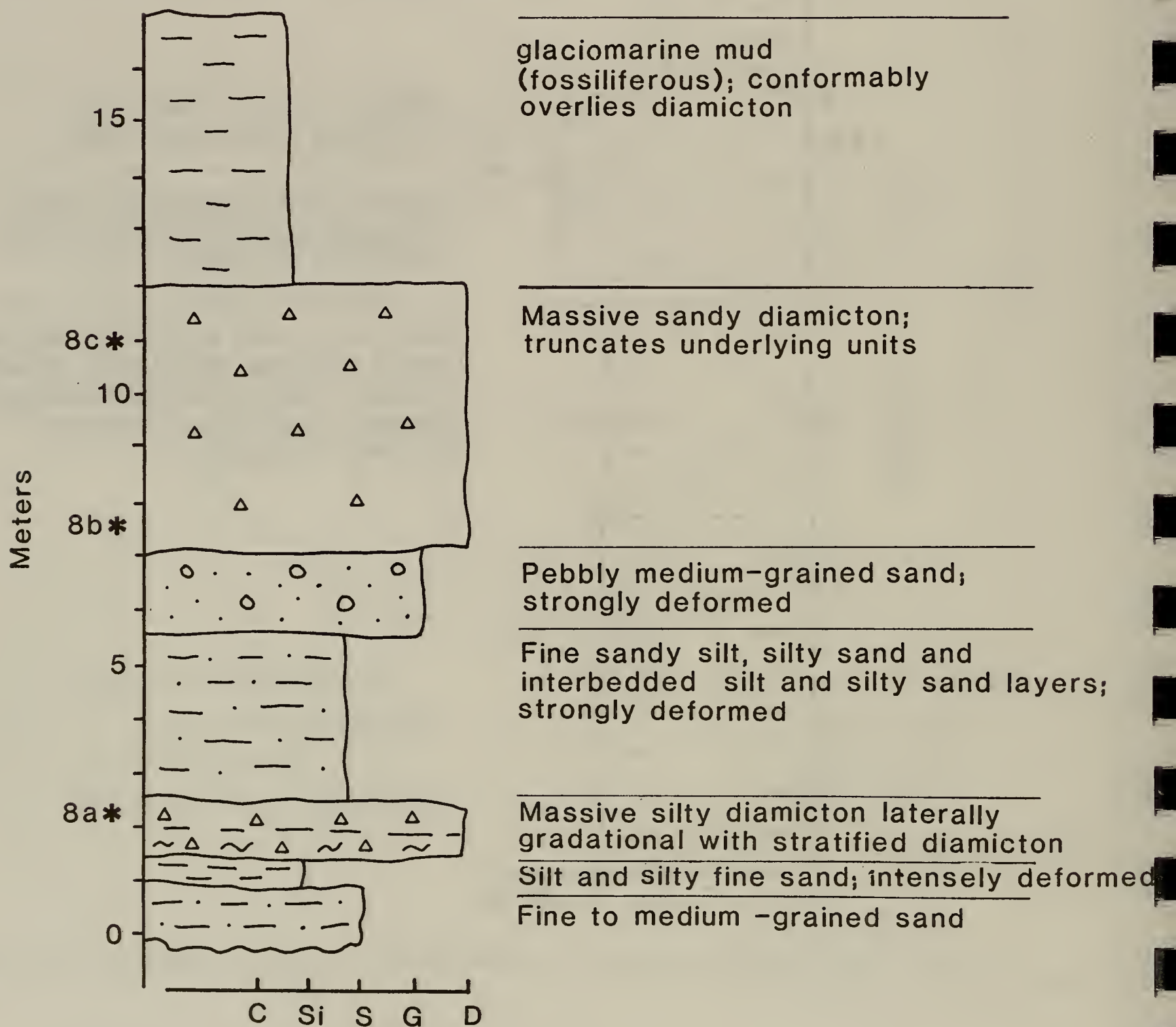


Figure 21. Schematic diagram of stratigraphy at Stop 8B (symbols same as Fig. 17).

The diamicton unconformably overlies 0.8 m of pale yellow to pale olive (5Y 7/3; 6/3) deformed, pebbly, medium-grained sand, which includes stringers and pods of diamicton. Beneath the sand, about 3.5 m of deformed pale olive (5Y 6/3; 6/4) fine sandy silt and silty fine sand is found. The base of the section is covered by slump.

Section 8B is found along the river bank approximately 200 m west of the access road section. Here, in general, 5 m of olive (5Y 4/3) glacimarine mud conformably overlies 5 m of pale olive to olive (5Y 6/3; 4/3) moderately compact sandy diamicton. This, in turn, unconformably overlies 1.5 m of deformed light yellowish brown (10YR 6/4) gravelly sand. Unconformably beneath the gravelly sand is 3 m of intensely deformed, layered light yellowish brown (10YR 6/4) to pale yellow (2.5Y 6/4; 7/4) to pale olive (5Y 6/3) fine sandy silt and silty sand. Underlying the deformed silt and sand is a 1 m thick olive to olive gray (5Y 4/4; 4/2) silty compact diamicton, which has a sheared, anastomosing platy fabric. This diamicton grades laterally into distinctly layered diamicton and silt beds, similar in appearance to the layered diamicton at Stop 7. Beneath the lower diamicton, deformed light gray to light olive gray (5Y 7/2; 6/2) silt and silty fine sand and light yellowish brown (2.5Y 6/4) to dark brown (7.5YR 3/4) fine- to medium-grained sand is present. Cross-ripple drift lamination is present in the fine-grained deposits, and planar-cross beds are found in the coarse-grained sand.

The sediments at Stop 8 record a similar history as that presented at Stop 7. The lower part of the section represents deposition of proglacial sediments along an advancing sublobe of ice from the Kennebec River valley, which was initially topographically controlled by the NE-SW trend of the Sandy River valley. The ice margin advanced into a proglacial lake where subaqueous depositional processes were occasionally interrupted by basal depositional and glacitectonic processes when the ice margin fluctuated. Later, the main phase of the Late Wisconsin ice overrode the early advance deposits laying down the upper sandy diamicton, which was subsequently mantled by the glacimarine mud (Presumpscot Formation) during deglaciation.

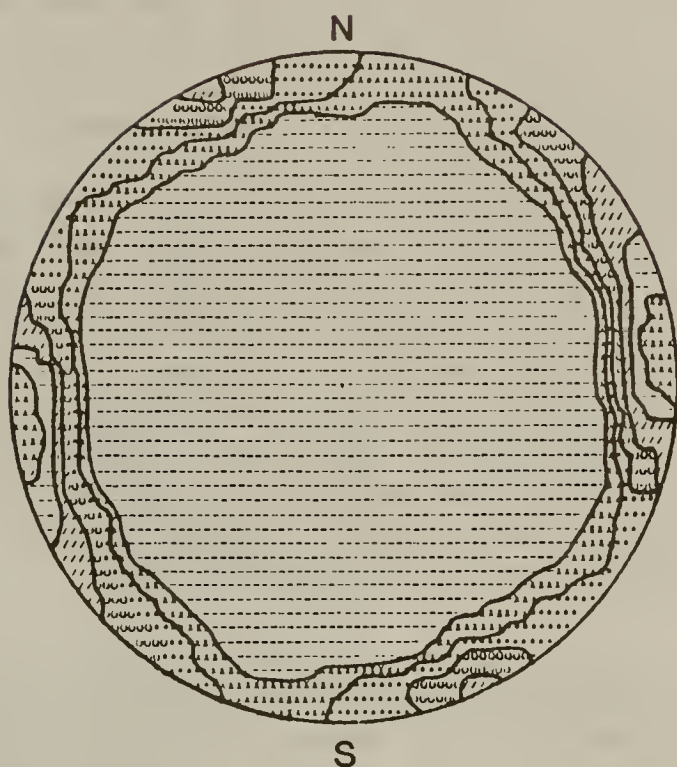
Petrofabric data from the two sections are shown on figure 22. Fabrics 8a, b, and c were reported by Richard Pratt (pers. commun., 1986) and are supplemented with additional data. The fabrics show a clear shift upsection in clast long-axis orientation, from a near E-W direction in 8a to a strong NW fabric in 8d. The strength of these fabrics ranges from 0.62 to 0.87, and bimodal distribution is present in the three fabrics from Stop 8B.

The glacitectonic data at Stop 8 (Figure 23) reflect the same upsection shift in structural components as is found at Stop 7, however, at Stop 8 this shift appears to be present in the petrofabrics as well as the structural data. There are two clear structural components in Figure 23; an early NE-trending set indicated by the slip-line analysis, bedding and thrust planes, and fold axial planes, and a later set represented by a cluster of thrust faults which have low to moderately-steep northwest-dipping surfaces.

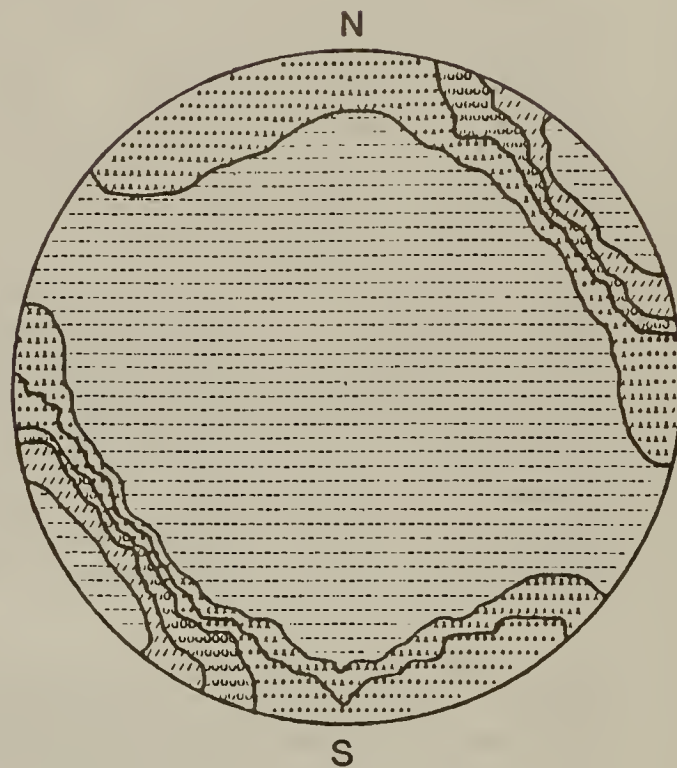
The upper sandy diamicton at Stop 8 is equivalent to the upper sandy diamicton at Stop 7, however, the upper unit at Stop 7 is not as coarse grained as the upper unit at Stop 8. This is clearly a function of the upper units at both stops being derived in part from the immediate substrate; at Stop 8 the upper diamicton overlies coarse-grained sediment, whereas at Stop 7



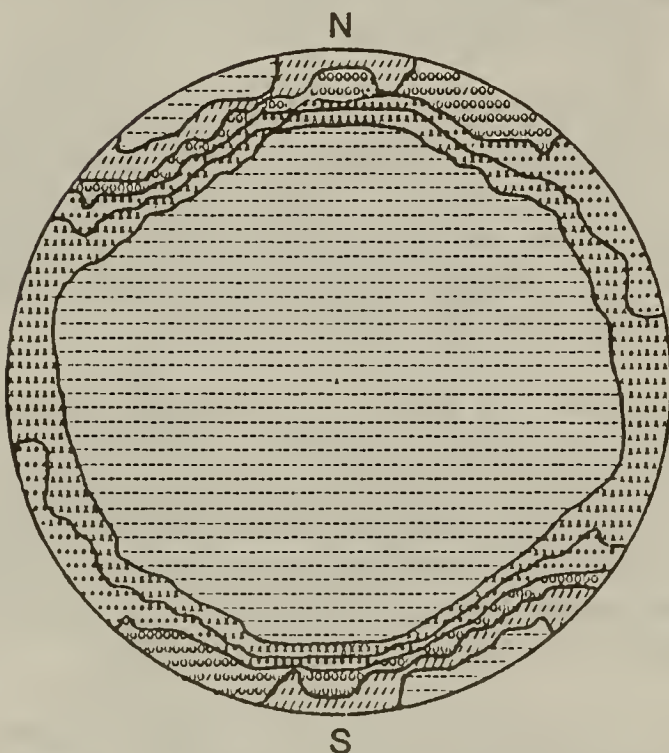
# Petrofabrics, Stop 8, New Sharon NEIGC, 1989



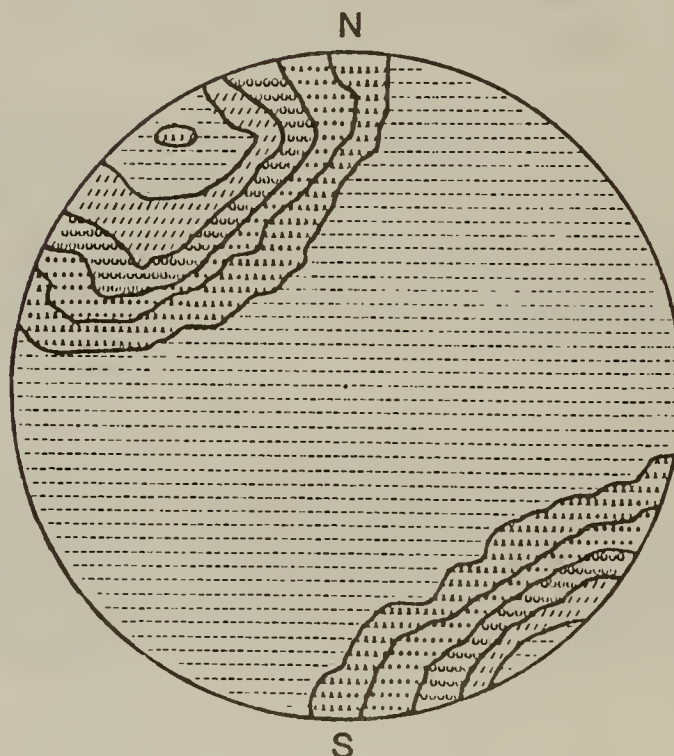
8a\*



8b\*



8c\*



8d\*

Figure 22. Petrofabrics from units at Stop 8; refer to Fig. 20 and Fig. 21 for locations (fabric data 8a to 8d respectively,  $n = 60, 47, 53, 25$ ;  $S_1 = 0.62, 0.67, 0.65, 0.87$ ;  $V_1 = 81, 0, 41, 0.5$ ;  $351, 0.5$ ;  $323, 10$ ; contour interval = 2 sigma).



#### Slip-Line Analysis Data

- ↷ Fold Axis (rotation sense noted)
- Slip Line (N18E39)
- Pole to Slip Plane (N56W40NE)

#### Planar Data

- Late Planar Features
  - Pole to Thrusts
- Early Planar Features
  - ◇ Pole to Thrusts
  - Pole to Bedding (parallel to early thrusts)

Figure 23. Glaciectonic data from units at Stop 8; two general fields are present, 1) early NE-dipping planar features and slip-line analysis from lowest units (below 7 m on Fig. 8B), and 2) later NW-dipping planar features from pebbly sand (below 3 m on Fig. 8A).



the upper diamicton overlies fine-grained sediment. The relation is even more obvious within each section, especially at Stop 8 where the lower silty diamicton overlies fine-grained deposits and the upper diamicton overlies coarse-grained deposits.

One point which may be argued at Stop 8 is whether or not the lower silty diamicton which is laterally gradational with layered diamicton is a good example of deformation till. Elson (1961, 1989) defined deformation till with the following characteristics; 1) it is derived from beneath the ice typically incorporating unconsolidated materials, 2) it is formed and deposited beneath the ice usually in an undrained basin, 3) has variable thickness, 4) range of deformation varies from total homogenization of primary sedimentary structures to displaced clasts having been moved only a short distance, 5) bedding may be present, especially in its upper part which may blend into massive till, 6) clast orientation should reflect deformation of a shearing matrix beneath the glacier, but a preferred orientation probably would be less well developed than fabrics in basal meltout or lodgement tills, 7) clast shape and surface marks are not diagnostic, 8) grain-size and lithologic characteristics depend entirely on the overridden substrate, although erratics from upgradient sources may be present 9), it commonly is "underconsolidated" and has low density due to the sediment having high pore water pressure during deformation, and 10) surface expression is seldom diagnostic.

This section was tentatively correlated by Caldwell (1959) with the exposure of the New Sharon Beds, in which the upper sandy diamicton was correlated with his Sandy River till (Late Wisconsinan age), and the deformed sand, silt, and lower silty diamicton was correlated with his New Sharon till (pre-Late Wisconsinan age). This tentative age assignment is important because although Locality C of Caldwell (1959; Figure 16, this report) is no longer exposed, the units which have been correlated to the two-tills of southern New England by Koteff and Pessl (1985) are present at Stop 8. The definitive criteria which prove the relative age of the upper till and lower till are not met at Stop 8, and thus that correlation is invalid. Furthermore, the clay mineralogy of the two diamicton units at Stop 8 indicates that both units have undergone clay mineral alteration but not to the degree which the lower till of southern New England has undergone (Newton, 1978; Weddle, in press). The stratigraphic and glaciectonic (kinetostratigraphic) data in association with the mineralogic data suggest that the section at Stop 8 represents advance and retreat of a glacier during a single glacial cycle, the Late Wisconsinan. The till at New Sharon is correlative with the upper till of southern New England. Evidence for Middle Wisconsinan, Early Wisconsinan, or Illinoian aged till is lacking, however, seismic data indicates that at least 24 m of high velocity seismic material (1800-2250 m/s) occurs beneath the New Sharon Beds. The age of this material is not known.

Finally, while the age of the deposits at New Sharon differs in this report from that originally proposed by Caldwell (1959, 1960), the interpretation that the deposits are in part the result of a glacier advancing into a lake was initially proposed by Caldwell (1959);

"....Evidently the Sandy River glacier advanced into a lake in which ...clay was deposited. The varved clay in the Sandy River till was formed during minor oscillations of the margin of the Sandy River

glacier. On the bedding plane surfaces of the varved clay layers well preserved lineations record the direction of movement of the overriding glacier. ...the New Sharon glacier, like the Sandy River glacier advanced into a body of standing water, and that clay was deposited over the New Sharon till following deglaciation, as happened following the melting of the Sandy River glacier."

199.1 0.5      Return to vehicles; leave Stop 8 and turn right at metal gate back to Route 2.

200.6 1.5      Stop; turn right onto Route 2 and proceed west to Farmington (approximately 10 miles).

END OF TRIP





## **HYDROGEOLOGY AND DEVELOPMENT PRESSURES BRUNSWICK, MAINE**

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### **INTRODUCTION**

This field trip is intended to focus upon the hydrogeologic aspects of a sand and gravel aquifer in Brunswick, Maine. It is developed from, and based upon, the results of a five-months long study done for the Town of Brunswick during the period October 1988 to March 1989. Conceptually structured by the Town's Planning Board and Conservation Commission, the study was intended to satisfy the conditions of a moratorium on residential development in a part of Brunswick referred to as the "Old Bath Road Area" (see Figure 1). It was enacted in Spring, 1988 by the Brunswick Town Council.

### **PREVIOUS WORK**

Wisconsin glaciation of the State of Maine and the subsequent geologic history of the area has resulted in the deposition of large volumes of sand and gravel. Many of these are aquifers. A significant aquifer, as defined by the Maine State Legislature (38 M.R.S.A., Chapter 3, Section 482 4-D), is a porous formation of sand and gravel that contains significant quantities of water that are likely to provide drinking water supplies.

Maine has a number of laws which restrict land use in various ways where such uses may have negative impacts on sand and gravel aquifers (see for example Site Location of Development Act, 38 M.R.S.A., Sections 481-490). In order to inform the State and local governments as they continue to develop laws and ordinances, the Maine Geological Survey and the United States Geological Survey began a joint mapping program in 1978 to identify, locate, and describe the existing sand and gravel aquifers in the State. This program, continuing until 1980, resulted in the publication of 59 maps summarizing the results of the investigation. These maps show approximate aquifer boundaries, estimates of potential well yields, and locations of some possible point sources of contamination of the aquifers. Subsequently, the State Legislature directed the Department of Environmental Protection and the Maine Geological Survey to update these maps, providing more information (38



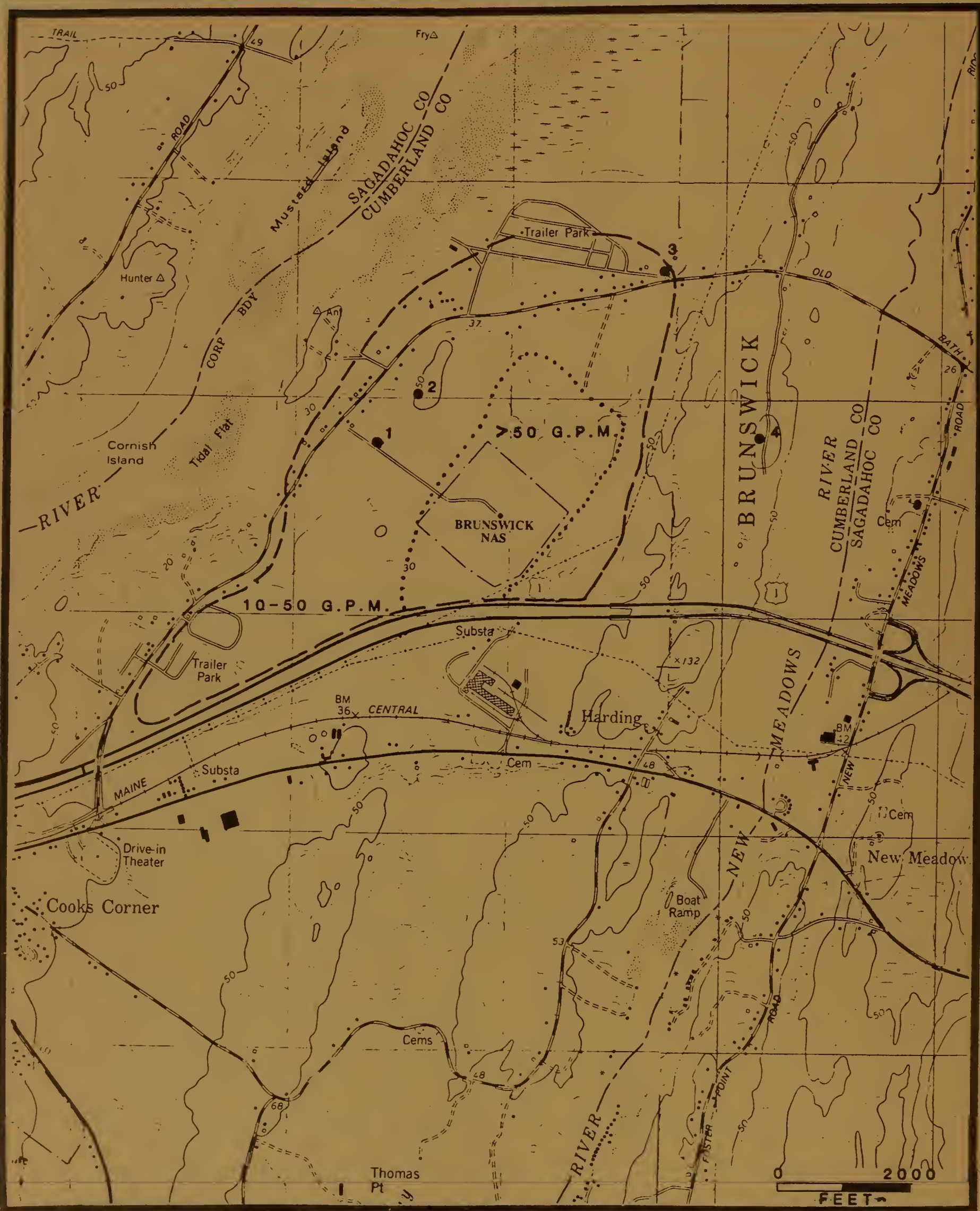


FIGURE 1 - LOCATION OF FIELD TRIP STOPS; PROJECTED WELL YIELDS



M.R.S.A., Chapter 3, Section 403). The additional work resulted in the publication of a new series of maps, referred to as Significant Sand and Gravel Aquifer Maps. Map #10 in that series, shows the aquifer in which the present study is located (See Figure 2).

Additional studies in the Brunswick area have been done by the E.C. Jordan Company for the Topsham/Brunswick Water District. These studies, focusing on the well fields which supply water to the District, include information relative to the relationship between the Androscoggin River and the sand and gravel aquifer.

## GEOLOGY OF THE STUDY AREA

### Bedrock Geology

As shown on the 1987 Bedrock Geologic Map of Maine and unpublished information of A.M. Hussey, II, the study area is underlain by various lithologies mapped as part of the Cape Elizabeth Formation. Few outcrops exist exposing these geologically old, and highly deformed, metamorphosed sedimentary rocks. The Cape Elizabeth Formation includes amphibolite, rusty weathering granofels, biotite schist, and quartzite. Some of these lithologies may be correlated with, and be the cause of, poor quality groundwater.

Mapping done to date has also confirmed the existence of geologically young, brittle fractures. These fractures, or fracture systems, have been named and in the area of the present study are represented by the Flying Point Fault and the Cape Elizabeth Fault (see Figure 2). Although, evidence of the existence of these structures is not directly observed, from relationships seen elsewhere the faults are inferred to be present beneath the surficial materials of the Old Bath Road area.

Prior to 2.5 million years ago, the metamorphosed rocks of the Cape Elizabeth Formation had been exposed and eroded to a gently rolling surface. An extensive mantle of soil existed, which had developed from the weathering of these metamorphic rocks. The topography then was presumably similar to present-day topography. It was characterized by northeast trending hills separated by long, narrow valleys. This is an expression of the "structural grain," which not only characterizes this part of Maine, but is also characteristic of much of the Appalachian Mountains of which this area is a part. The metamorphic rocks have been deformed into a series of folds which trend northeast-southwest and the folds in the bedrock control the topography which has evolved as the surface has been molded by the prolonged weathering and erosion of the rocks.

### Surficial Geology

The bedrock surface was traversed by great thicknesses of ice, beginning approximately 2.5 million years before present. Ice moved very slowly across the landscape. During alternating, advance and retreat, sediment entrained within it was deposited. Material deposited directly from ice, which is referred to as "till," contained a wide variety of particle sizes and showed no evidence of layering (stratification). Material which was deposited by meltwater streams, derived from the back-wasting of the ice sheet, was stratified due to aqueous transport and is referred to as "outwash."



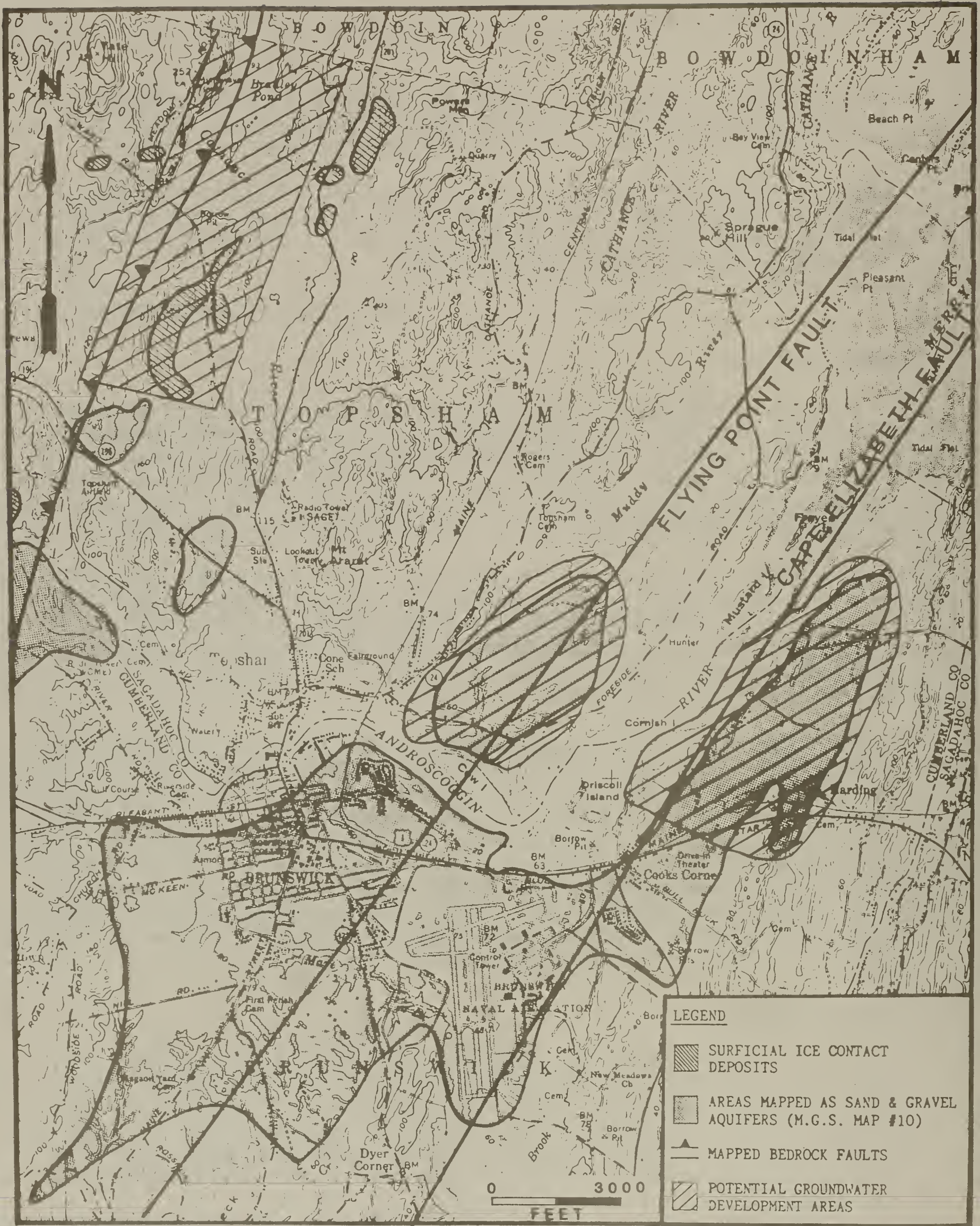


FIGURE 2 - POTENTIAL GROUNDWATER SUPPLY LOCATIONS  
(MODIFIED FROM E.C. JORDAN CO., 1986)



The study area is a large outwash plain characterized by fine to coarse sands and fine gravel. It represents slightly more seaward deposition of sediment than the nearer-shore deposits of deltaic coarse sand and gravel. Examples of the latter exist to the north across the river in Topsham.

Outwash deposits, consisting of deltaic sands and gravels and more distal sandy deposits, were deposited contemporaneously with, and succeeded by, younger accumulations of marine silt and clay as sea level rose and transgressed the land surface. The deltaic and outwash deposits were, in part, eroded by near-shore currents during the deposition of these younger and finer marine sediments.

The crust of the earth, free of the weight of the ice sheet, rebounded. Sea level (relative to the land) began to fall. As the shoreline retreated seaward, much of the previously deposited outwash and marine sediments were reworked by the streams which drained the land. As a result, stratified sediments have been deposited in places above the fine-grained marine silts and clays. In the study area, these near-surface coarse sands and fine gravels are typically very poorly stratified. They imply braided streams which had constantly changing channels as they eroded and re-worked the previously deposited sediments. Although subtle, it can be seen that some of the surface topography is due to stream erosion as the "ancestral" Androscoggin River shifted its channel throughout this area.

To obtain more information about the subsurface stratigraphy, the logs of 5 previous borings and samples from 5 new borings were examined. The borings totaled 147.5 feet. Most borings intersected very poorly sorted, and hence only weakly stratified, medium sand to sandy gravel at shallower depths. At greater depths, the sediments became finer and distinctly stratified. Some of the holes, but not all, encountered blue-gray marine clay and clayey silt.

The stratigraphy shows significant variation. Borings to the east all intersect sands which are fine to coarse, as well as some gravel. Four MDOT test borings along U.S. Route 1, although they are separated by only 100 feet, appear to reflect a transition between stratigraphy with significant clay layers to the west, and stratigraphy devoid of such fine-grained material to the east.

The reasons for the differing stratigraphy in the eastern and western parts of the study area is not clear. The two areas are separated by a till mantled bedrock ridge, which extends from the benchmark located on the north side of U.S. Route 1 northeast to the south side of the Old Bath Road. This bedrock ridge may have controlled currents in the near-shore area which existed during the time of sediment deposition. Such currents could have removed the fine fraction of deposited sediment to the east while the absence of such currents west of the bedrock high may be the reason why a much larger percentage of fine material has accumulated there.



## AQUIFER CHARACTERISTICS

### Measurement of Water Levels

Because the depth to unsaturated material is nowhere greater than 15 feet, and in many places is less than 3 feet, residents of the Old Bath Road area rely on dug wells, or driven points, for domestic water. Thus, at numerous locations the ability to monitor groundwater level, measure hydraulic conductivity of the aquifer, and retrieve samples for chemical analysis already existed. Monitoring wells, constructed of 2" diameter PVC, installed in 6" hollow-stem auger borings, and 1 1/4" diameter stainless steel well points were established to supplement these existing observation points. In summary, the data points were as follows:

1. 2" observation wells (5)
2. 1 1/4" diameter stainless steel well points (5)
3. large diameter concrete-lined or stone-lined dug wells (5)
4. pre-existing observation wells (6)
5. ponds (2)

### Estimates of Permeability

Permeability was calculated or estimated by three methods. Falling-head and constant-head field test data were used to calculate permeability in a manner suggested by Hvorslev (1951). Permeability was also estimated from the lithological descriptions of sediment encountered by borings in the study area. This was done using data presented in Weiss et al (1982). The values determined using the latter approach were confirmed by reference to Masch and Denny (1966).

### Estimates of Yield

Yields to domestic wells in the aquifer, as well as to the five 2" monitoring wells installed in the course of the study, were estimated using a method proposed by Mazzaferro (1980). The yield in gallons per minute was determined by multiplying the transmissivity by the aquifer thickness and dividing by 750. This method has been used by the Maine Geological Survey in the evaluation of sand and gravel aquifers.

These estimates, coupled with field observations on the distribution of sediments, identified an approximate area which would yield in excess of 50 gallons per minute to properly installed wells. It lies within the larger area of the sand and gravel aquifer, where it is estimated that 10 gallons per minute is an expectable yield (see Figure 1).

Two features of the outwash aquifer are especially noteworthy. The first is that the sand aquifer is not very thick (though silty sands are in places up to 60 feet thick). Thus, though permeabilities in sandy portions of the aquifer are high, transmissivities (permeability multiplied by aquifer thickness) are not. Consequently, few locations within the study area offer the potential for high volume wells necessary for municipal water supply. Secondly, many parts of the outwash aquifer are laminated sand and silt. In such portions of the aquifer, horizontal permeability greatly exceeds vertical permeability.

## GROUNDWATER QUALITY

The large number of dug wells and driven points attest to the fact that groundwater quality at present in the Old Bath Road area is very high. However, continued growth and development in the Town of Brunswick threatens to impact water quality. The following are either existing or potential sources of contaminants, which in the future could significantly change water quality. Each is briefly discussed.

### Septic Systems

Septic systems have been in use in rural areas in the United States and elsewhere for many years. Despite the fact that design requirements have been changed from time to time and the systems presumably improved, many studies indicate unacceptable impact on groundwater quality from contaminants introduced with wastewater. All single-family homes in the Old Bath Road area presently dispose of their wastewater through individual septic systems.

Figure 3 illustrates the behavior of septic system contamination plumes in sand and gravel aquifers. The threat to shallow water supply wells is clear.

### Underground Storage of Petroleum Products

Underground storage of petroleum products is a serious form of groundwater contamination in Maine (Garrett, 1986). But in the study area, there are only two registered underground tanks, both used for heating oil in the vicinity of the offices of H.C. Crooker & Sons, Inc.

### Agriculture

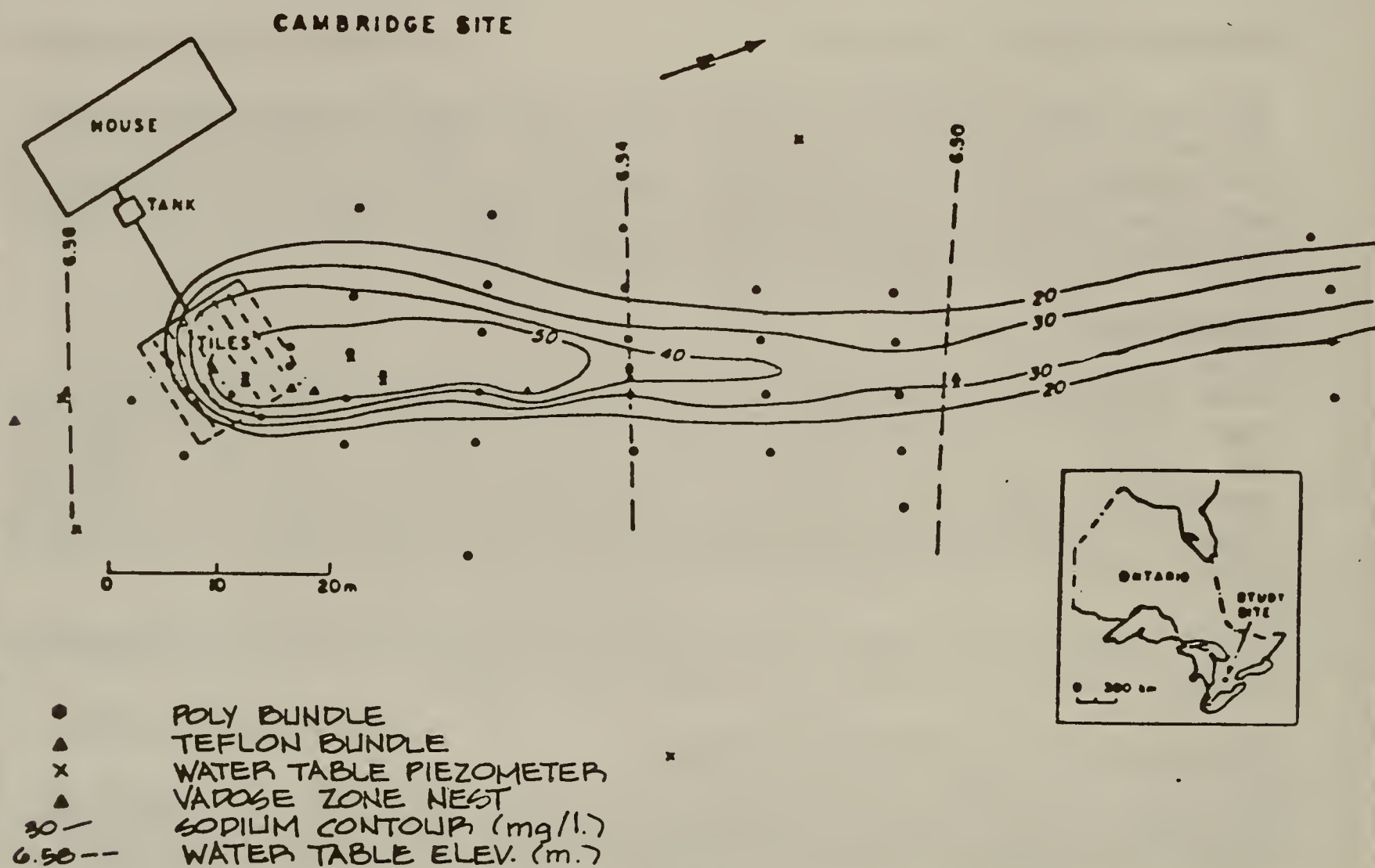
There is currently limited agricultural activity in the area of the aquifer. It includes:

1. horse boarding and pasturing (17 horses)
2. a greenhouse: Paul's Produce & Greenhouses
3. rental garden plots: Stewarts Gardens
4. two small dairy farms

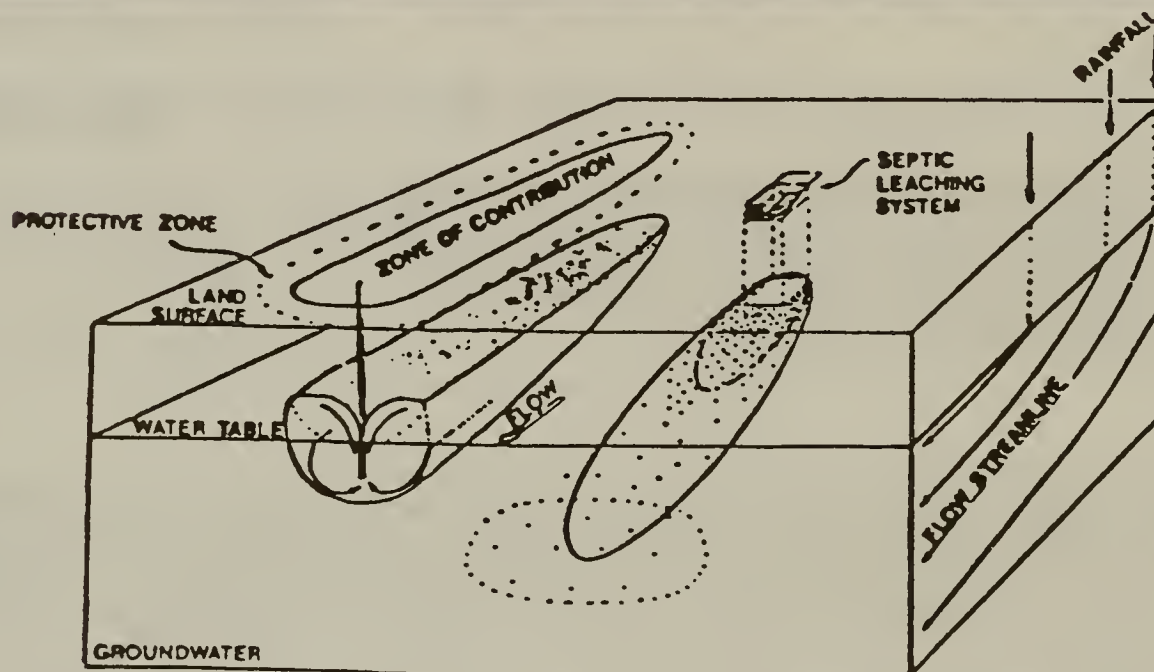
### Road Salt

At the present time, the U.S. Environmental Protection Agency and the Maine drinking water standard for sodium is 20 mg/l for public supplies and 100 mg/l for private wells. There is ongoing consideration of the defensibility of these standards and in the near future it can be expected that the standard may be changed or eliminated. Because of this, consideration of sodium chloride contamination of groundwater is subject to some debate. However, the Maine Department of Transportation uses approximately 3.75 tons of salt per lane mile per year on State-maintained roads. Therefore, the 2.3 miles of Route 1, which traverse the boundary of the aquifer study area, represents a significant source of salt introduction (34.1 tons). Records from the Town of Brunswick suggest an effort to reduce the amount of salt which is spread on the roads maintained by the Town. The Town used 11.4 tons per mile in the winter of 1986-87. Direct observation of the Old Bath Road indicates use of significantly less salt than is used on the average throughout the Town.





(AFTER ROBERTSON ET AL, 1989)



(AFTER KERFOOT, 1987)

FIGURE 3 - CONTAMINANT PLUMES IN SAND & GRAVEL AQUIFERS

Therefore, it is very difficult to estimate the input of sodium chloride to groundwater as a consequence of road maintenance.

### Sea Water

The contamination of domestic wells by salt water intrusion, as well as by "fossil sea water," in the State of Maine have both been documented. Some bedrock wells that have been drilled through surficial materials and into the underlying bedrock have elevated concentrations of sodium chloride. The reason for these high concentrations is not known. However, the presence of "fossil sea water" in the fractures in bedrock is a suspected source.

Where water is obtained from the bedrock, used and disposed of in a septic system, the effluent sodium and chloride concentrations can be expected to be augmented by the fact that the source water was also higher than usual in concentrations of these two elements. The real threat to sand and gravel aquifer water quality posed by the discharge to it of water from bedrock wells is obviously highly speculative.

## SUMMARY

The information presented above, in addition to the brief field trip which follows, are both intended to provide sufficient information for discussions of the impact of development upon the resources of the study area. For the information of trip participants, page 66 of the Town of Brunswick Zoning Ordinance is reproduced here...

### 404 Aquifer Protection Zone

404.1 Definition of Zone. The Aquifer Protection Zone consists of sand and gravel aquifers and primary and secondary aquifer recharge areas, as identified in the report "Hydrologic Evaluations for Designation of Aquifer Protection Zones, E.C. Jordan Co., January 1986".

404.2 Special Use Provisions for Sand and Gravel Aquifers and Aquifer Recharge Areas. The following special use provisions are applicable to:

- A. Prohibited Uses. The following uses are prohibited.
  - (1) The disposal of solid waste other than brush or stumps;
  - (2) The disposal or storage of hazardous matter, as defined in Section 201.32;
  - (3) The disposal of leachable materials, except these from Dwellings, Single and Two-family;
  - (4) The storage of road salt or other de-icing agents.
- B. Special Exception Uses. The following uses may be permitted by special exception:
  - (1) All uses permitted in primary districts except those noted above;
  - (2) Animal feed lots;
  - (3) Manure piles and storage pits;
  - (4) The spreading of chemical fertilizers;
  - (5) The storage of petroleum or other refined petroleum



- products, except petroleum stored on residential property, to be used for residential purposes;
  - (6) Aerial spraying of pesticides;
  - (7) Piling or storing bark.
- C. Special Exception Conditions. In addition to the requirements in Section 705, the uses permitted by special exception are also subject to the following:
- (1) The use must not deplete ground water supplies;
  - (2) It must not interfere with aquifer recharge;
  - (3) It must not lower the quality of potable ground water;
  - (4) The following information must be provided to the Zoning Board of Appeals to assist it in making a decision:
    - (i) The immediate and long-range impact on the ground water;
    - (ii) The amount and types of waste to be granted by the proposed activity, and the adequacy of the disposal system;
    - (iii) The topography and drainage of the site and its susceptibility to flooding.

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Tepper, D.H., Williams, J.S., Tolman, A.L., Prescott, G.C. Jr., 1985, Hydrogeology and water quality of significant sand and gravel aquifers in parts of Androscoggin, Cumberland, Franklin, Kennebec, Lincoln, Oxford, Sagadahoc, and Somerset Counties, Maine: Maine Geological Survey Report Open-File No.85-82a, 106 p.

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### ITINERARY

MEETING PLACE: large parking lot on north side of Zayre Store, located on the south side of Old Route 1, 0.5 miles east of Cook's Corner, Brunswick, Maine.

#### Mileage

- 0.00 intersection of Old Bath Road and Old Route 1
- 0.15 overpass over Route 1...Maplewood Mobile Home Park is served by a sewer extension which passes over this bridge and joins a portion of the collection system of the Brunswick Sewer District.
- 0.40 four-legged nitrate sources in pasture south of road
- 0.50 Maplewood Mobile Home Park, located on the south side of the Old Bath Road
- 0.75 small detention pond designed in accordance with storm water run-off calculations (25 yr./24 hr. storm)...Site Location of Development Law, ch. 375, section 4 B.1.
- 1.40 turn right on paved access road to the Communications Building, Naval Air Station, Brunswick

STOP 1: We will leave the cars at this location and begin a traverse across the sand and gravel aquifer beginning near Route 1 and walking north across the Old Bath Road to the vicinity of the Androscoggin River. On this traverse, we will observe the upper portion of the sand and gravel deposits as revealed in gravel pits (now ponds). In addition, we will conduct falling head and constant head field permeability tests. The rental garden plots referred to above, can be observed. Finally, we will look at seeps which represent the discharge of groundwater to the Androscoggin River. After the traverse is completed, return to

- (1.40) vehicles, turn around and proceed back to the Old Bath Road, turn right



1.60 turn left, turn around, and park

STOP 2: We will leave the vehicles for another short walk. The purpose here is to get an overview of a proposed development, focusing on the very difficult questions of appropriate design of water supply and wastewater systems. Return (1.60) to cars, turn left, and proceed in an easterly direction.

2.50 entrance to Bay Bridge Estates, turn left.

STOP 3: We will simply drive slowly through the development, allowing observation of the development density and wastewater disposal facilities. The latter will not be examined in detail, but some of the component parts of the system can be (2.50) observed. Return to Old Bath Road and turn left.

2.65 intersection of power line with Old Bath Road

2.85 Bridge Road intersection, turn right.

3.40 STOP 4: The purpose of this stop will be to look at the western boundary of the sand and gravel aquifer, noting the very different character of the topography

#### END OF FIELD TRIP

**THE CHAIN LAKES MASSIF  
AND ITS CONTACT WITH A CAMBRIAN  
OPHIOLITE AND A CARADOCIAN GRANITE**

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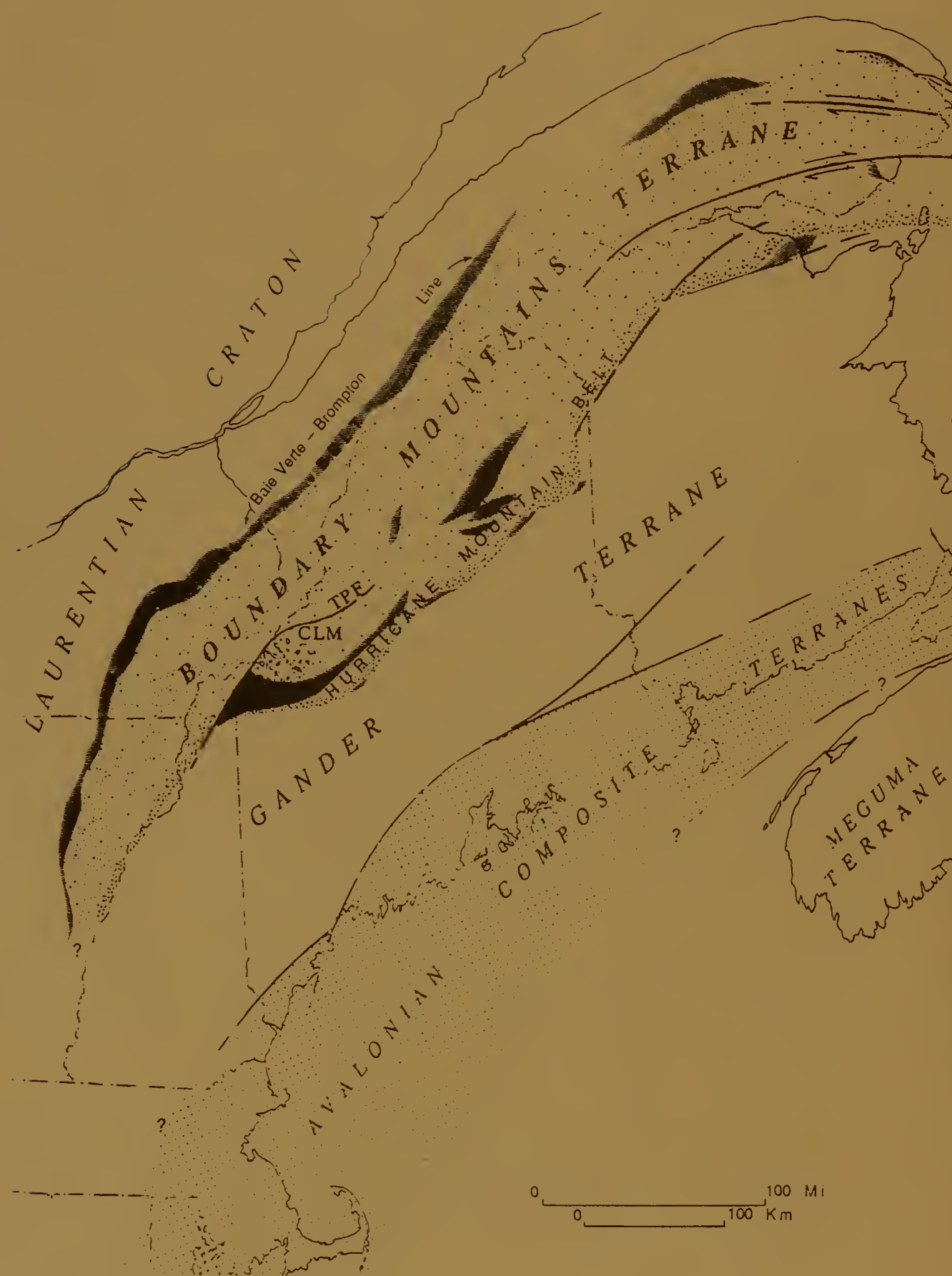
**INTRODUCTION**

The Boundary Mountains along the Maine — Québec border form the height of land and watershed between drainage northward to the St. Lawrence River and southeastward to the New England part of the Atlantic coast. Along this part of the international boundary the highest peaks attain elevations of 3500 to 4000 feet (~1070-1220 m), and are held up by resistant rocks of the Chain Lakes Massif and the western part of the Boil Mountain ophiolitic complex (Boudette, 1982). The Boundary Mountains constitute the massif and the principal exposure of this distinctive basement, but exotic blocks of Chain Lakes granofels are found in the St. Daniel mélange in southeastern Québec and in its equivalent, the Mélange ophiolitique du Ruisseau Nadeau, in the southeastern part of Gaspé peninsula (de Broucker, 1987), as detailed in Boone and Boudette (1989), and references therein. Inasmuch as the type locality of the Chain Lakes granofelsic basement is in the Boundary Mountains, we have termed the tectonic terrane underlain by sialic basement that is bounded by the two ophiolite and mélange belts — the St. Daniel, along the Baie Verte Brompton line, and the Hurricane Mountain belt to the southeast — the Boundary Mountains terrane (fig. 1). The Boundary Mountains terrane was progressively covered by forearc turbiditic flysch of Late Cambrian to Early Ordovician age, followed by Mid- to Late Ordovician felsic and mafic volcanics and an unconformable northwest-transgressive shoreline-shelf, cyclic turbidite succession beginning in the Silurian. These rocks range in age up to Lower Devonian, immediately preceding the Acadian orogeny. Rocks of the Boundary Mountains terrane are intruded by calc-alkalic granitic rocks. Their ages range from Late Ordovician to Middle Devonian (fig. 2; table 1).

The principal purpose of this trip is, first, to examine several lithologic and structural variants of Precambrian metamorphosed diamictite constituting perhaps the major rock type as seen in outcrops of the Chain Lakes Massif (Boudette and Boone, 1976), and second, to examine a part of the Hurricane Mountain Belt, where (1) the Boil Mountain ophiolitic complex is combined within a tectonostratigraphic succession with (2) metamorphosed volcanics and volcanogenic metasedimentary rocks (Jim Pond Fm.) and mélange (Hurricane Mountain sequence), both probably of Middle to Late Cambrian age (table 1), and (3) a forearc basin flysch sequence (Dead River Fm.) of Late Cambrian to Early Ordovician age, which overlies the mélange. The volcanics - mélange - flysch succession is almost entirely unfossiliferous.

The Hurricane Mountain mélange is separated from the Chain Lakes Massif (CLM) by the Boil Mountain complex (Boudette, 1982). The Boil Mountain ophiolite lies in tectonic contact with the Chain Lakes Massif. The fault pattern (figs. 2a and 2b) is complicated by numerous normal and transverse faults, but the initial juxtaposition is best interpreted as





**FIGURE 1.** Index map showing the accretionary configuration of major lithotectonic terranes in the northern Appalachian orogen. CLM, Chain Lakes massif; TPF, Thrasher Peaks fault. Modified from Boone and Boudette (1988, Fig. 4A).

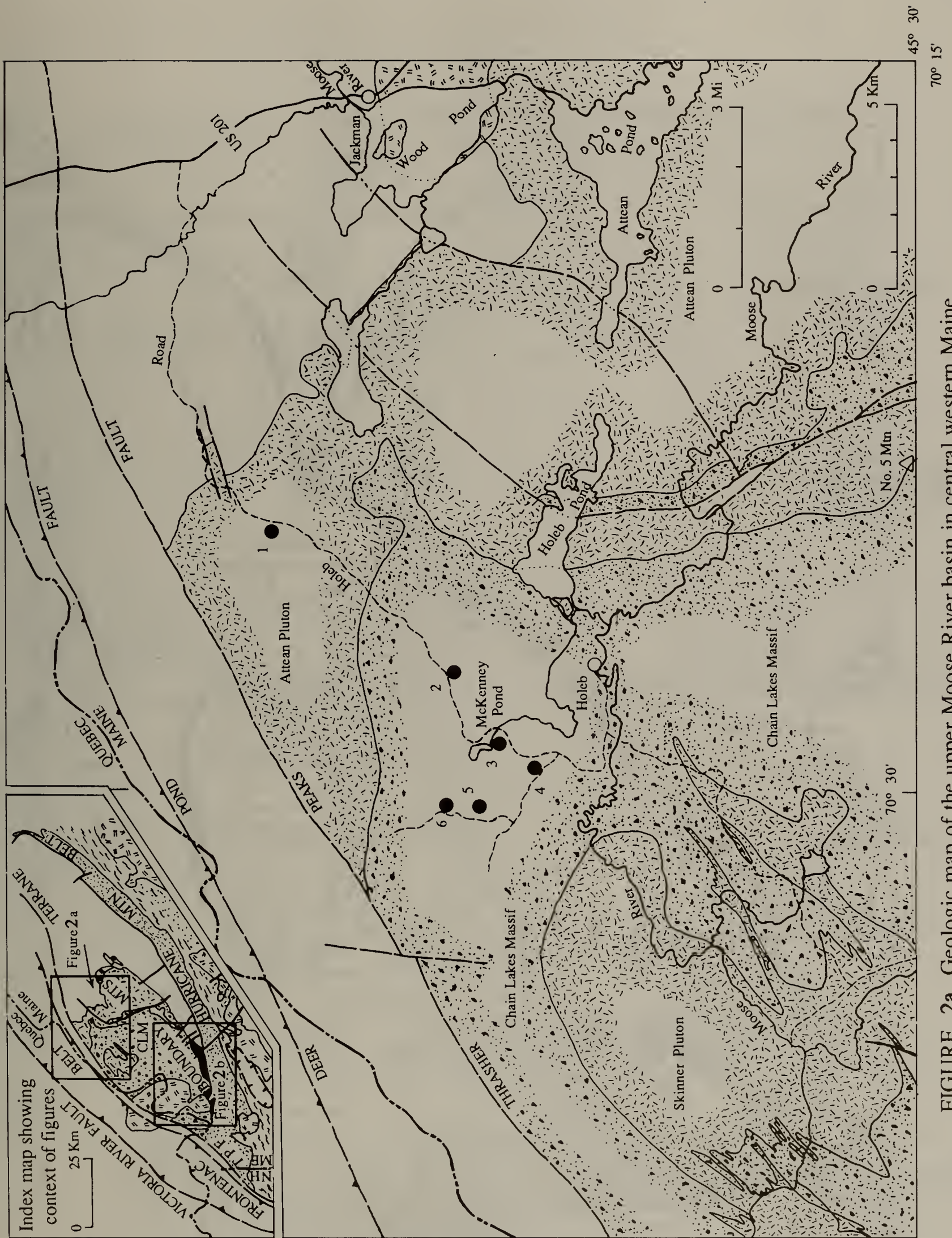


FIGURE 2a. Geologic map of the upper Moose River basin in central western Maine.





FIGURE 2b. Geologic map of a highway transect from Coburn Gore to Eustis in central western Maine.

# EXPLANATION

## METAMORPHOSED SEDIMENTARY ROCKS, VOLCANIC ROCKS, AND SUEVITE(?)

## INTRUSIVE ROCKS

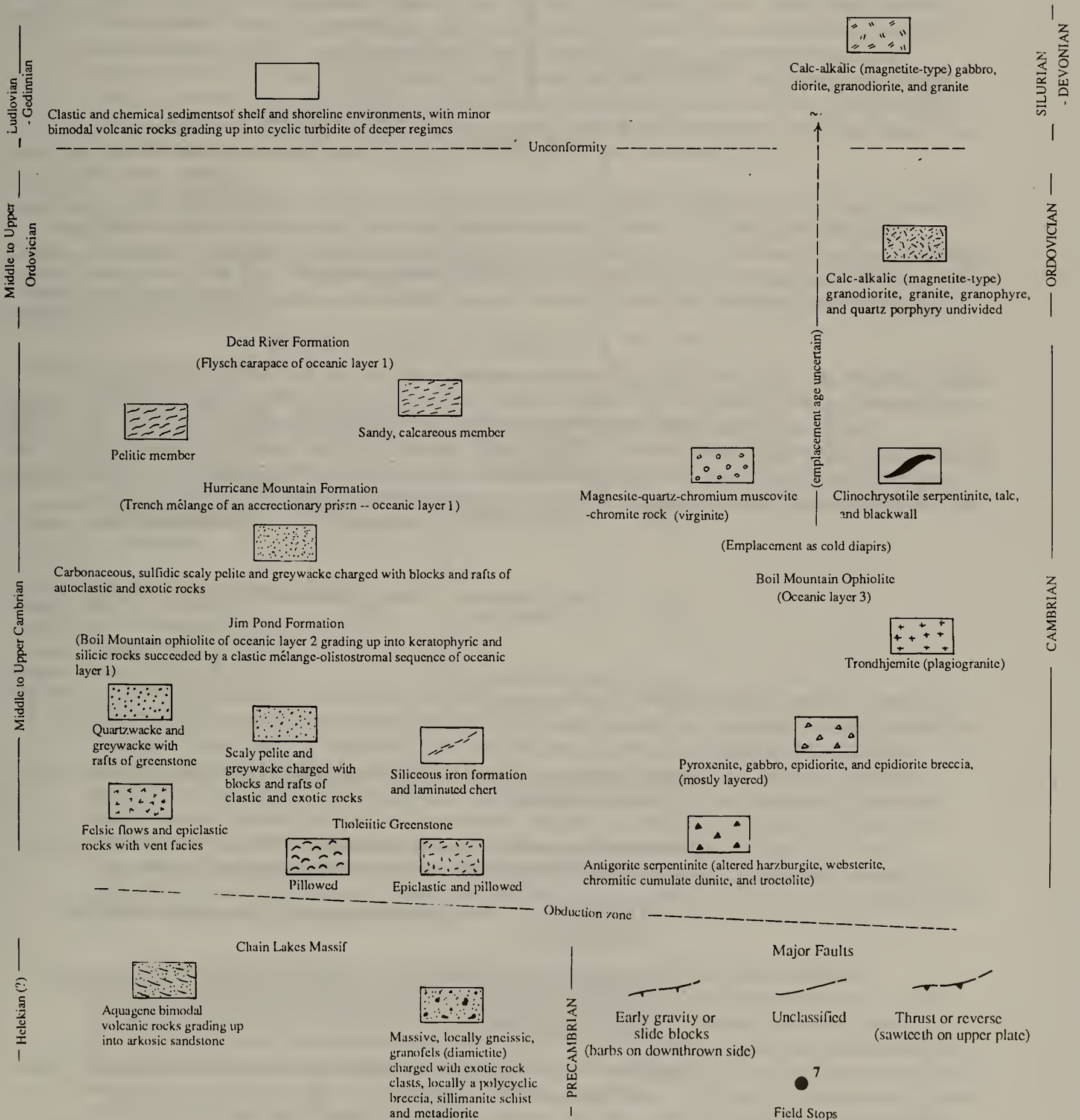




Table 1. Chronology of tectonometamorphic events in the region containing the Chain Lakes massif.

ISOTOPIC AGE* (Ma)	TIME-STRATIGRAPHIC INTERVAL	EVENT	INTERPRETATION
367.6 ± 1.3	Middle Devonian	Intrusion of Lac des Araignées Granite	Prograde contact metamorphism produced aureoles <1 km wide; high-T assemblage: Bio + Staur + sillimanite = garnet
373.3 ± 2.0		Intrusion of Chain of Ponds Pluton	
367.7 ± 1.3		Intrusion of Big Island Pond Pluton	
397 ± 10	Early Devonian	post-Acadian intrusion of Lexington Batholith	Transposed foliation, SE margin of Chain Lakes Massif
418 ± 14		Acadian Orogeny	
443 ± 4	Late Ordovician	Intrusion of Attean batholith	Prograde contact metamorphism, ca. 200 - 300 m wide aureole, biotite + sillimanite = garnet hematite-bearing assemblages
ca. 455 - 475	NO EVIDENCE OF TACONIAN OROGENY		
520 ± 12	Late Cambrian		Post-obduction plagiogranite, Boil Mtn Complex
540 > t > 484	Late Cambrian (extending to Early Ordovician ?)	Penobscottian Orogeny	Obduction of Boil Mtn ophiolitic Complex; accretion of Hurricane Mtn. mélange; thrust and fold tectonics of accretionary complex & forearc basin deposits
ca. 570	Early Cambrian	Rift to drift, passive margin	Cooling through Ar blocking -T of hornblende, Chain Lakes granofels
ca. 770	Late Proterozoic (Hadrnynian)	Crustal thinning & rifting	High T/P thermal metamorphism development of fleck gneiss in volcanogenic granofels
1534 ± 60	Mid-Proterozoic (Helikian)	?	Age of Chain Lakes protolith: formation of diamictite; impact metamorphism in northern part of Chain Lakes Massif ?

Middle Devonian  $^{40}\text{Ar}/^{39}\text{Ar}$  cooling ages of hornblende granitic plutons for Heizler, Lux, and Decker (1988).

Minimum age of Acadian orogeny in western Maine (Rb/Sr mineral and whole rock) from Gaudette and Boone (1985), and Boone and Gaudette (unpub. data)

ca. 418 Ma (Rb/Sr) age of retrograde metamorphism of Chain Lakes granofels, from Cheatham (1985).

U/Pb zircon age of Attean batholith, from Lyons and others (1986). U/Pb zircon age of Boil Mtn. plagiogranite, from Eisenberg (1982) and John N. Aleinikoff (pers. commun., 1988).

Preliminary  $^{40}\text{Ar}/^{39}\text{Ar}$  age bracket (amphiboles in meta-basites) of Penobscottian orogeny, from Boone and others (in press).

$^{40}\text{Ar}/^{39}\text{Ar}$  cooling age of hornblende, Chain Lakes granofels from Biederman (1984).

Rb/Sr average age of high T/low P metamorphism of Chain Lakes rocks, from Cheatham (1985).

U/Pb zircon age of Chain Lakes granofels protolith, Naylor and others, 1973.

obduction of the ophiolite along a consuming plate boundary. In the area shown by fig. 2 and elsewhere along strike to the northeast, we interpret the mafic and ultramafic rock fragments in the *mélange* as derived from obduction and fragmentation of oceanic crust, (presumably early Paleozoic Iapetus oceanic lithosphere), along a suture zone between adjacent suspect accreted terranes. The Boundary Mountains Terrane, lying northwest of this suture zone, is floored by the mainly sialic Chain Lakes granofels and gneiss, and stands within the orthotectonic part of the northern Appalachians of Maine and Québec. It lies between the southeasternmost surface exposures of Laurentian (Grenvillian) crust, in southeastern Québec, and the northwesternmost exposures of Avalonian basement, southeast of the Kearsarge - Central Maine synclinorium (KCMS; see Lyons and others, 1982). Beneath the overlapping Late Ordovician to Early Devonian strata of the KCMS, a third suspect terrane, inferred to be the Gander (Boone and Boudette, 1989) separates the Boundary Mountains Terrane from those identified as Avalonian (fig. 1).

The deformation within the Hurricane Mountain belt is inferred to have been caused by suture zone collision resulting in the amalgamation of two sial-floored terranes: the Boundary Mountains and Gander. This early accretionary event is termed the Penobscottian orogeny (after the "Penobscottian disturbance" of Neuman; see Neuman and Rankin, 1966). It is separated in space, and probably largely in time, from the Taconian orogeny which followed it (table 1).

The second part of the trip, along and near State Hwy. 27, lies along the western Maine seismic reflection/refraction profile (see Stewart, 1989, for a summary and an interpretation of the data). Extrapolation of the surface geology of the Chain Lakes massif and surrounding tectonostratigraphic units down dip of reflectors along this profile is convincing in the interpretation that the Boundary Mountains terrane, with its Chain Lakes basement, was thrust upon the Grenvillian basement of the early Cambrian continental margin, in this western Maine part of the northern Appalachians.

## CHAIN LAKES MASSIF

### Regional Context

The CLM (fig. 1) is bounded by faults and contacts with intrusive rocks. Along its northwestern margin, slate-grade Devonian strata are juxtaposed against the CLM along the Thrasher Peaks fault. To the north and east of the transect, the massif is bounded by intrusive contact with the Late Ordovician Attean batholith (table 1). Locally along its eastern margin, the CLM is overlain unconformably by late Silurian and early Devonian strata. Along its southeastern and southern margins, the massif is fault-bounded by the ophiolitic rocks of the Boil Mountain Complex, and by metafelsite of the Jim Pond Formation. The strike of these units and the fault which separates them diverges sharply from the regional northeasterly trend, to northwesterly, around the southwest margin of the CLM in the southwest (figs. 1 and 2b).

Approximately 3,000 m of exposed structural thickness is estimated for the polymetamorphosed granofels, gneiss, and minor amphibolite and metasandstone of the CLM. This estimation is derived from an interpretation that the overall structure is an elongated foliation dome with principal plunge to the southwest, along the regional strike (Boudette and Boone, 1976). According to this interpretation the structurally lowest rocks are in the northeastern part of the massif. There, the rocks, which may lie structurally above well layered, quartzofeldspathic to pelitic gneiss, are characterized by a relict fragmental fabric which led Boudette and Boone (1982) to propose an exometamorphic, impact origin to account for a suevite-like, relict micro- and meso-structure, prior to regional



metamorphism. Rocks containing this fabric are the subject of the first part of this field trip. The structurally highest 1000 m, in the southern part of the massif, is composed of layered metavolcanic and interspersed meta-arkosic rocks.

Much of the massif (~2000 m), as exposed in the valley of the Chain of Ponds and North Branch of the Dead River transected by State Highway 27 (fig. 2b), is characterized by massive to weakly layered granofels of approximately dacitic composition. Gneissic variants are common locally. Variations in composition are marked by modest changes in the ratios of minerals comprising the stable assemblage of biotite, chlorite, muscovite, plagioclase, epidote, alkali feldspar and quartz. Different ratios of alkali feldspar and plagioclase are the most notable. Magnetite and hematite are common oxide accessories. Relict almandine-rich garnet, cordierite and sillimanite (fibrolite), occurring singly or in pairs with the stable greenschist assemblage in semipelitic rocks, can be found in many localities scattered throughout the massif. Relict AFM phases, however, have not yielded consistent or meaningful thermobarometric results. The polymetamorphic history of the CLM, moreover, spans a period of 800 Ma or more and records a generalized PTt path of prograde and retrograde events terminating in the Devonian.

The most distinctive feature of most rocks of the massif is the occurrence of relict boulders and pebbly fragments distributed sparsely to commonly throughout the granofelsic or gneissic matrix, and readily visible in most outcrops. Their sizes range from a few centimeters to as much as a meter in long dimension, but most are 2 to 15 cm in length. Thus much of the massif consists of polymict, originally matrix-supported diamictite. Many of the fragments are of foliated, semipelitic and semicalcic schists and fine-grained, thinly laminated gneiss. Randomized orientations of their foliations with respect to layering or foliation of host granofels or gneiss indicates derivation from source areas of pre-existing metamorphic rock. Granitic rock, diorite, gabbro and felsic volcanic rock clasts are dispersed throughout the diamictite, but nowhere known to be abundant. Less common fragments that are also foliated, consist of banded amphibolite. Some may be exotic, but others appear to be tectonic inclusions which originated by boudinage of thin amphibolite layers followed by separation and rotation of the boudinaged fragments. Most of the fragments of presumed exotic origin are commonly rounded, but may also be angular and blocky. Well-rounded clasts are commonly ellipsoidal with long axes paralleling their internal structures but not necessarily parallel to the layering or foliation of the enclosing matrix. Rotation of clasts, apparently synmetamorphic, is common. At the hand-specimen and thin section scales, angular to rounded nodules of quartz, with or without cores of epidote and sulfide minerals, are quite common. Their origin (clasts vs. metamorphic segregations) has often provoked debate at the outcrop. Amphibolite is a sparse lithologic component of the structurally highest parts of the massif, where gneissic structure, parallel to compositional layering, is more common.

The structurally (and probably stratigraphically) lowest sequence of the CLM is subdivided into three facies: (1) the Twin Bridges semi-pelitic gneiss; (2) the Appleton epidiorite; and (3) the Barrett Brook polycyclic epidiorite breccia. We propose that this sub-diamictite sequence comprises the vestiges of a target-rock succession if the impact hypothesis (Boudette and Boone, 1982) is valid. The principal diamictite sequence is presently subdivided into four distinct facies, each with internal variants of structure, texture, and clastic components. These facies are: (1) the McKenney Pond chaotic rheomorphic granofels; (2) the Coburn Gore semi-pelitic gneissic granofels; (3) the Kibby Mountain flecky gneiss, and (4) the Sarampus Falls massive to layered granofels. The structurally (and probably stratigraphically) highest sequence is the Bag Pond Mountain bimodal metavolcanic section and feldspathic meta-arenite. The sequence is well-layered and upright as indicated by depositional structures. The Bag Pond Mountain rocks are



interpreted to represent a return to sedimentation and volcanism in a passive margin or epicontinental setting following impact metamorphism and deformation.

## Structure, Petrology, and Genesis

Biederman (1984, p. 32 - 53) has contributed a uniquely relevant and detailed discussion of the petrography and structure of the Chain Lakes rocks along Route #27. Within the transect of the trip, the Chain Lakes is disposed as a broad arch in which foliation and bedding parallel. The rocks are typically dark gray to gray-green and medium- to coarse-grained. Layering, clast population and the distribution of metamorphic minerals provide the only distinction between lithic types.

The rocks along the transect are composed principally of quartz, sericitized plagioclase, chlorite and chloritized biotite, muscovite, and K-feldspar. Subordinate sillimanite (fibrolite), garnet, apatite, allanite, sphene, zircon, epidote, calcite, and opaque minerals (hematite, magnetite and sulfides are found) (table 2). These minerals are arranged in a variety of granoblastic textures. Biederman (1984, p. 38) recognized two distinct modal assemblages attributable to metamorphic overprint. One assemblage (sillimanite + biotite + quartz + plagioclase + K-feldspar + garnet) is attributable to a high-grade event. The second assemblage (quartz + muscovite + epidote + chlorite + calcite) is interpreted to be related to Paleozoic low-grade events.

One of the distinctive lithic types to be seen in the Chain Lakes is flecky gneiss (Stop 3) in which fibrolite-oxide-rich clots, which characterize the rock, appear to have been produced by the breakdown of cordierite or garnet. Another (Stops 6 and 9) is the tectonite produced along the boundary with the Boil Mountain ophiolite in the Sarampus Falls facies.

## Metamorphism

The CLM is characterized throughout by chlorite + muscovite + brown or green biotite-bearing assemblages that overprint relict high amphibolite to low (biotite-bearing) granulite facies assemblages imposed upon thickly bedded to massive protoliths, most, to varying degrees, diamictic. The relict high grade assemblages constitute massive granofelsic to well-foliated gneissic meso- and microstructures. Intermediate between these structural extremes are relict flecky gneisses, localized patches of agmatitic rock, and breccia, in which a relict foliation is variously poorly to well developed. The present chlorite-biotite grade assemblages are for the most part mimetic after these earlier-formed structures.

Within the central part of the massif, through which the route of the field trip lies, the varied degrees of fleck-gneiss abundance and structure are impressive. This is probably the reason for Rankin and others (1983) reference to the rocks of the massif as "gneiss and migmatite of Chain Lakes" in the correlation chart for Precambrian rocks of the Eastern United States. Our view, gained over several years of mapping within the massif, is that a massif-wide application of the term migmatite is misleading.

The polymetamorphic history of the CLM spans a period of 800 m.y. or more and records a pressure-temperature-time path of prograde and retrograde events terminating with the emplacement of the Late Ordovician Attean batholith (table 1). We believe that the high-grade regional metamorphism(s) took place in the Proterozoic. It is doubtful that the high-grade aspects of the massif could be ascribed to Grenville tectonometamorphic events. Even if an early phase of the high-T metamorphism was coeval with the Grenville rocks, it was followed by a period of relatively static, widespread high temperature production of



Table 2. Chemical and modal analyses of rocks from the Chain Lakes massif and Boil Mountain ophiolite.

Major oxide composition (weight percent)									
Chain Lakes massif <sup>(1)</sup>							Boil Mountain ophiolite <sup>(2)</sup>		
Sample No.	1	2	3	4	5	6	7	8	9
Field No.	CL-38	CL-40	CL-42	CL-34	S-38	BP-13	784a	115a	267a
SiO <sub>2</sub>	67.2	69.9	69.2	68.8	68.4	65.8	43.8	50.7	34.0
Al <sub>2</sub> O <sub>3</sub>	15.0	14.7	15.3	13.7	15.4	16.7	15.9	15.7	14.8
Fe <sub>2</sub> O <sub>3</sub>	2.8	2.3	1.3	3.5	1.3	2.2	3.6	1.5	2.3
FeO	2.9	2.4	4.0	1.9	3.2	3.5	7.3	5.1	3.3
MgO	1.6	1.2	1.3	1.3	1.2	1.7	10.4	10.1	32.5
CaO	1.5	1.8	1.3	1.5	1.6	1.6	15.6	11.7	0.8
Na <sub>2</sub> O	2.7	2.6	1.5	2.5	2.6	2.0	0.18	2.2	0.04
K <sub>2</sub> O	2.4	3.1	3.1	2.9	2.6	2.9	0.14	0.46	0.02
H <sub>2</sub> O <sup>+</sup>	1.9	1.4	2.5	2.9	2.5	2.5	2.1	2.5	12.1
H <sub>2</sub> O <sup>-</sup>	0.07	0.01	0.04	0.01	0.01	0.07	0.02	0.06	0.35
TiO <sub>2</sub>	1.0	0.90	0.81	1.0	0.78	0.93	0.21	0.36	0.24
P <sub>2</sub> O <sub>5</sub>	0.17	0.18	0.13	0.19	0.16	0.16	0.01	0.04	0.01
MnO	0.09	0.07	0.08	0.07	0.08	0.10	0.12	0.10	0.02
CO <sub>2</sub>	0.41	0.26	0.10	0.27	0.03	0.10	0.02	0.04	0.02
Total	99.74	99.82	100.66	99.14	99.66	100.26	99	101	101
Normative mineral composition (weight percent)							CIPW norm		
q	38.3	37.5	43.0	40.2	37.9	37.2	0	0	0
c	6.7	4.9	7.8	4.9	6.0	8.1	0	0	15.2
or	14.8	18.6	18.7	17.6	15.8	17.6	0.85	2.78	0.1
ab	23.3	22.3	12.9	21.6	22.6	17.3	1.6	19.0	0.4
an	3.8	6.2	5.1	4.7	6.9	6.5	43.5	32.3	4.3
en	4.1	3.0	3.3	3.3	3.1	4.2	cp 29.6	21.4	0
fs	1.6	1.2	5.2	0.0	3.8	3.3	op 2.0	14.6	22.3
mt	4.1	3.4	1.9	3.5	1.9	3.3	ol 22.0	9.0	57.3
hm	0.0	0.0	0.0	1.1	0.0	0.0			
il	1.9	1.7	1.5	1.9	1.5	1.8	0.4	0.7	0.5
ap	0.4	0.4	0.3	0.0	0.4	0.4	0.02	0.1	0.03
ca	0.9	0.6	0.0	0.6	0.1	0.2	0.1	0.1	0.1
Total	99.9	99.8	99.7	99.5	99.8	100.0	100.7	100.0	100.1
DI	87	89	87.5	89	89	87	2	24	1

(continued)

Table 2 (continued)

Modal analyses									
Points									
counted	1234	1040	1052	1083	1056	1155			
Sample No.	1	2	3	4	5	6	7	8	9
Quartz	33	38	46	42	42	36	tr	x	--
K-feldspar	1	0.4	0.6	0.5	5	2	--	--	--
Plagioclase	37	28	15	34	22	24	--	x	--
Biotite	9	7	6	0	0	0	--	--	--
Muscovite	12	21	22	17	19	23	--	--	--
Chlorite	5	1	10	4	10	13	(20)	x	(5)
Epidote	1	1	0.1	0	0.3	0.3	(35)	x	--
Iron oxide	1	2	0.6	2	1	1	(5)	x	(5)
Apatite	0.3	0.4	0.2	0.1	0.3	0.1	--	--	--
Calcite	0	1	0.1	0.6	tr	0.2	--	--	--
Zircon	tr	tr	0	tr	tr	0	--	--	--
Allanite	tr	0	0	0	0	0	--	--	--
Sillimanite	tr	0	0	0	0	0	--	--	--
Leucoxene	tr	0.4	0	0.4	0.2	tr	--	--	--
Sphene	0	0	0	0	0	tr	--	--	--
Amphibole?	0	0	0	0	0	0	--	--	--
						Actinolite	--	x	--
						Green hornblende	(45)	x	--
						Clinochysotile & Mg-chlorite	--	--	(90)

(1) Rapid rock analysis, H. Smith, 1984

(2) Rapid rock analysis, H. Smith, 1969-70

- Not observed

tr Trace

x Present in major amounts

(35) Thin section estimate

1. Sarampus Falls facies, Chain Lakes massif; gray, plicated gneissic diamictite containing short discontinuous streaks of mica and scattered clasts of quartz in quartzo-feldspathic matrix. Outcrops on Rte. 27, 2.1 km NW of junction with Beaudry Road, Chain of Ponds quadrangle. Lat. 45° 19' 30", Long. 70° 38' 45".
2. Kibby Mountain facies, Chain Lakes massif; gray, faintly gneissic diamictite containing scattered clasts of quartz of different sizes and elongate lenses in which clusters of biotite as much as a centimeter in diameter give a spotted appearance to the rock, matrix is quartzo-feldspathic. Chain of Ponds quadrangle, Rte. 27, 0.9 km NW of junction with Baudry Road. Lat. 45° 19' 10", Long. 70° 37' 50".
3. Bug Eye Pond facies, Chain Lakes massif; dark gray, massive diamictite containing different size clasts of quartz and biotitic metamorphic rock scattered in a quartzo-feldspathic matrix. Road cut Rte. 27, 600 m NW of Sarampus Falls, Jim Pond quadrangle. Lat. 45° 18' 20", Long. 70° 36' 50".

(continued)



Table 2 (continued)

4. Bag Pond facies, Chain Lakes massif; light gray, white-weathering, even-grained, fine-grained, faintly pin-striped gneiss. Quartz clasts rare. Outcrops on trail at 1800 ft contour, 800 m N of Round Mountain Pond, Chain of Ponds quadrangle. Lat. 45° 16' 45", Long. 70° 38' 50".
5. Bug Eye Pond facies, Chain Lakes massif; gray, massive diamictite containing scattered clasts of quartz of different sizes and numerous mafic-rich clasts as much as 1 cm in diameter in quartzo-feldspathic matrix. Ledges at bridge over Bog Brook, Skinner quadrangle, Lat. 45° 31' 30", Long. 70° 32' 25".
6. Bug Eye Pond facies, Chain Lakes massif; dark gray, massive diamictite containing scattered clasts of quartz and mafic-rich rock in quartzo-feldspathic matrix. Cuts along railroad, 1.7 km NE of Lowelltown, Boundary Pond quadrangle, Lat. 45° 31' 20", Long. 70° 37' 35".
7. Troctolite, massive, equigranular with bladed hornblende up to 3 mm long, western Toe Nail Ridge, Black Mountain quadrangle, Lat. 45° 14' 49", Long. 70° 40' 20".
8. Gabbro, massive, medium- to coarse-grained intergrown plagioclase and ferromagnesian minerals producing a diabasic to graphic texture, east of Greenbush Pond on east side of North Branch, Dead River, Tim Mountain quadrangle, Lat. 45° 14' 59", Long. 70° 30' 19".
9. Serpentine, diapiric type, waxy, sheared with cross-fiber veins up to 1 cm wide, along south Tea Pond road, Tim Mountain quadrangle, Lat. 45° 14' 06", Long. 70° 31' 26".

In CLM rocks most plagioclase is altered all or in part to sericite; muscovite includes both sericite not recognizably from alteration of plagioclase and discrete flakes; chlorite and iron oxide derived primarily from alteration of biotite; some sericite in Samples 1 and 2 from former fibrolite.

granofels, flecky granofels, and flecky gneiss, prior to the earliest Phanerozoic tectonothermal events.

Exposures of the contact to be visited (Stops 5 and 6), preserve a thin tectonic lens of amphibolite at the base of the Boil Mountain Complex. The amphibolite shows a range of texture of the constituent amphiboles, ranging up to coarse-grained patches of hornblende prisms, which define the foliation that parallels the contact. Other matrix minerals are sphene, epidote, opaque oxide(s), and sparse chlorite, which are fine-grained. Chlorite is confined to small patches, presumably having replaced actinolite or hornblende. The amphiboles show a range of composition. The largest grains are euhedral to subhedral, very fresh hornblende, showing here and there a sharp zonation to, or mantling by actinolite at their margins. Actinolite also occurs as small single grains and clusters amid medium- to fine-grained hornblende lenses adjacent to the patches of coarse-grained hornblende.

Oft-cited in favor of obduction-related metamorphism is the observation of coarser-grained muscovite in the common quartzofeldspathic rocks of the Chain Lakes along its southeastern and southern perimeter at Stop 6 (Boudette, 1970, 1978, and 1982). Although this hypothesis has not been investigated or tested with respect to microstructural change, it still stands as viable. One consequence of this possibility is the requirement of H<sub>2</sub>O mobility at elevated temperature within an inferred structural thickness of about half a kilometer below the obduction contact. Coarse muscovite, however, is not everywhere present within this zone.

## THE BOIL MOUNTAIN COMPLEX AND JIM POND FORMATION

Boudette (1982, p. 212) and Coish and Rogers (1987; p. 51) have described the illogical practice of dividing the plutonic from the volcanic and sedimentary components of the ophiolite. For descriptive purposes alone, the ophiolite-mélange-flysch carapace succession retains herein the subdivision used by Boudette (1970, 1978, and 1982) and Boudette and Boone (1976), and Boone and Boudette (1989). Coish and Rogers (1987) have established the geochemical context of the Boil Mountain Complex (Boudette, 1982) combined with the volcanic components of the Jim Pond Formation. According to Coish and Rogers (1987) the Boil Mountain ophiolitic complex is probably best correlated in the Appalachians with the Dunnage Zone ophiolite of Newfoundland. They also furnish support for realistic comparisons of the Boil Mountain with other world-wide ophiolite complexes.

### Boil Mountain Complex

Lithologic units include serpentinite, pyroxenite, gabbro, trondhjemite, and epidiorite of oceanic layer 3 (see table 2). Rocks of the Boil Mountain Complex are extensively altered, but relatively pristine examples may be found in the northeast. Deformation and dislocation of the rocks of the complex increase toward the southwest along with alteration, especially in the serpentinite. This effect becomes especially notable where faults on the southeast become tangential to the ophiolite. Extreme tectonism and metamorphism is believed to have remobilized segments of the serpentinite of the complex to produce the diapiric variety. The serpentinite of the Boil Mountain Complex contains antigorite and is moderately hydrated; diapiric serpentinite described in a later section, in contrast, is more hydrated. Antigorite and clinochrysotile do not coexist in these rocks.



The distribution of rock types within the Boil Mountain Complex is relatively uneven and, to some degree, they are mixed. The two-fold subdivision of the ultramafic and mafic components shown on figure 2b is generalized and reflects the dominant rock types in each. The northeastern part of the complex has a stratigraphy wherein the rocks are part of a sequence as much as 1.6 km thick facing southeast and are relatively enriched in Mg at their base along the northwestern margin. The lower zone is represented by the principal occurrence of the antigorite serpentinite (altered harzburgite and dunite) and pyroxenite. The Mg-rich rocks (Stops 5 and 6) are generally overlain by massive epidiorite which in turn grades into gabbro, epidiorite autobreccia (Stop 10; epidiorite and subordinate trondhjemite), and minor clinopyroxenite. Greenstone septa (or possibly dikes) are common in the trondhjemite facies (Stop 7). Repetition of lithologies is common near the base, where distinctive igneous layering is also seen (Stops 5 and 6).

The base of the Boil Mountain Complex is exposed at Stops 5 and 6 where epidiorite, gabbro, or serpentinite of the ophiolite are in sharp, tectonic contact with rocks of the Chain Lakes massif. The latter appear as discrete septa within the ophiolitic rocks and show subtle effects of thermal recrystallization. Thus the basal contact is difficult to characterize at the scale of outcrops, but some protoclasia of the ophiolitic rocks suggest that ductile faulting accompanied their emplacement and is the pre-eminent relationship. Moderately well-foliated amphibolite shows partial to extensive retrogradation of hornblende to actinolite. Elsewhere along the base of the complex, the contrast in mechanical competence between the rocks of the Chain Lakes massif and those of the Boil Mountain Complex has resulted in the localization of post-intrusion fault dislocations (Stop 6) characterized by brittle deformation, and relationships are usually obscure.

### **Diapiric serpentinite, soapstone, and virginitite**

Discordant bodies of dominantly strongly-sheared serpentinite (see #10, table 2) associated with subordinate, but variable amounts of soapstone and virginitite (carbonate-quartz-chromium muscovite-chromite rock) intrude rocks of the Jim Pond Formation along faults (Stop 12). The diapiric serpentinite is notably hydrated and is composed of clinochrysotile and magnesian chlorite. The diapiric serpentinite has apparently detached itself from the parent complex and has migrated as much as 7 km from source areas.

Exposures of the bodies are nowhere sufficient to observe any consistent geometric arrangement of the serpentinite, talc rock, and virginitite. In most cases, the arrangement appears to be unique with each of the rocks successively in contact with country rock which is strongly sheared. In most cases where all lithologies are present, the talc rock tends to envelop the serpentinite and the virginitite envelops the other rocks. Both serpentinite and virginitite have been found to occur singly.

### **Jim Pond Formation**

The Jim Pond Formation is a combined bimodal volcanic, and aquagene volcanic - olistostromal sequence representing oceanic layers 1 and 2. On the northwest the Jim Pond is composed essentially of chlorite-albite-epidote-actinolite greenstone (Stop 8) with minor amounts of mafic-rich metagraywacke, metamorphosed dacite, maroon phyllite, and hematitic chert (jasper).

In the southeast the greenstone constitutes all but about 150 m of the Jim Pond section in the east, and gives way by short-ranged facies change to metaquartzwacke toward the



west. Greenstone is present as slide blocks within the metaquartzwacke member. The greenstone is thickly layered with lenses throughout characterized by uniformly- and well-developed pillows. The thick units are provably individual flows that are 15 m or more thick. Mafic lapillite, in layers 1-20 cm thick, and volcanic breccia compose less than 10 per cent of the greenstone member and are found throughout interlayered with pillowed and massive flows.

In addition to metamorphosed dacite, the Jim Pond Formation contains sodic quartz-latite flow rock and related ash-flow rock, breccia, and epiclastic rock (Stop 11). The thickness of the metadacite member varies from 0 to more than 500 m. The metadacite occurs in layers that are about 15 m thick or more. Ash-flow deposits are finely laminated in beds 1 to 10 mm thick. The boundary between the greenstone of the mafic member and the metadacite is sharp. Repetitive sequences of the two volcanic rocks occur with individual flows averaging about 30 m in thickness. The metadacite and iron-formation members are almost everywhere closely associated. In the northeast on regional strike, the main belt of metadacite is succeeded (toward the southeast) by iron formation with interlayered metaquartzwacke and metagraywacke.

## HURRICANE MOUNTAIN FORMATION

The Hurricane Mountain *mélange* (Boone, 1989, and references therein) represents part of an accreted wedge of carbonaceous, sulfidic scaly metapelite and metasiltstone which is charged with blocks and rafts of autoclastic and exotic rocks. If the Hurricane is accreted within a zone of orthogonal subduction, high-pressure mineral assemblages apparently were not formed. Localized occurrences of different exotic lithologies along the strike of the Hurricane suggest that subduction may have been oblique, and that concomitant strike-slip faulting, within the forearc environment or arcward of it, brought different provenances into the zone of active fragmentation where gravity-driven submarine slides were incorporated into the growing accretionary wedge.

The Hurricane retains a rather consistent structural thickness of 900 to 1000 m throughout the Lobster Mountain anticlinorium; this thickness probably is largely a product of Penobscottian, rather than Acadian deformation. The structural relationship of the Hurricane to the underlying, less deformed, aquagene volcanic Jim Pond Formation is essentially a fault contact, involving break-up and olistostromal emplacement of Jim Pond greenwacke, quartzite and volcanogenic rocks in a matrix which is increasingly composed of siltstone protolith structurally upward into the Hurricane in the southwest part of the Hurricane Mountain Belt (Fig 1; Boudette, 1978). We recognize the base of the Hurricane here as defined by matrix that is predominantly metasiltstone, commonly rusty, owing to disseminated pyrite and pyrrhotite (Stop 14). Farther along strike to the northeast, the Hurricane is in sharp structural discontinuity with felsic volcanics, or with graded beds of wacke and volcanogenic, bedded pseudochert and lenses of ferromanganese oxide of the upper part (estimated to be the upper one-third) of the Jim Pond Formation. A predominant axial planar cleavage in metasedimentary and metavolcaniclastic rocks of the Jim Pond is locally folded and weakly overprinted by a second, presumably Acadian cleavage.

## DEAD RIVER FORMATION

The Dead River Formation (Boone, 1973) is interpreted to have accumulated as a coeval flysh carapace over the Hurricane Mountain *mélange* seaward from the accretionary margin. The sediments of the Dead River were apparently transported into place by turbidity currents, and also reworked thereafter by bottom currents. An upward increasing



abundance of poorly-sorted metasandstone beds composed both of immature and unstable detritus in the Dead River indicates a gradual change into a higher energy sedimentologic environment.

The Dead River is subdivided into a lower metapelite member and an upper metasandstone member (Stop 13). The metapelite member consists dominantly of green, red, and variegated red and green phyllite and slate with minor amounts of calcareous metagraywacke in beds 5-10 cm thick. A few thin metalimestone lenses occur at the base of the member. The lower member is transitional into the upper member with the amount of metagraywacke and thickness of beds gradually increasing upward. The upper half contains up to 50 percent or more of metagraywacke and arkosic metasandstone, with metaquartzwacke beds ranging from 2 to 30 cm thick. The balance of the lithology is similar to the metapelites of the lower member. The contact between the lower and upper member is arbitrarily taken to be the horizon where the metasandstone beds comprise at least 50 percent of the lithology.

The rocks of the Dead River are associated in a variety of depositional structures ranging from parallel lamination to wavy and flaser-bedding to thickly bedded, graded sets. In general, parallel bedding and uninterrupted graded bedding are more common in the lower half of the unit, becoming less common in the upper half because of the increase in abundance of zones of small-scale chaotic structure, convolute structure, and other well-preserved soft-sediment deformational features. Despite the wide variety of bedding, a characteristic feature of metapelites throughout the unit is a pinstriped appearance produced by quartz- and quartz-feldspar-rich laminae parallel to cleavage and relict bedding. Metagraywacke and metaquartzwacke beds also commonly contain quartz-rich laminae developed along fracture-cleavage and slip-cleavage surfaces. These pin-stripe structures persist through a wide range of metamorphic conditions. The average regional stratigraphic thickness of the Dead River present is estimated to be 760 m (Boone, 1973), but as much as 1200 m could be present in places.

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## TRIP A-4 and B-4

### (ROCKS OF THE CHAIN LAKES MASSIF AND BOIL MOUNTAIN COMPLEX)

This two-day trip will traverse a combination of federal and state highways and privately maintained lumbering roads. The latter present appropriate hazards, but are navigable by ordinary vehicles in wet or dry weather. USGS quadrangle maps that may be helpful include the following:

#### Day #1

Attean	15'
Spencer Lake	15'
Skinner NE	7 1/2'
Skinner SE	7 1/2'

#### Day #2

Chain -of -Ponds	7 1/2'
Jim Ponds	7 1/2'
Spencer Lake	15'
Stratton	15'

The De Lorme Maine Atlas and Gazetteer (De Lorme Publishing Company, Freeport, ME - - available at most Maine bookstores and sports outlets) is also recommended. Background geology for the trip is provided in the references.

Day #1 will be directed toward roadside outcrops in the Moose River headwaters in the vicinity of Holeb and mainly concerned with the rocks of the northern part of the Chain Lakes massif. Woodland access, always at the discretion of nature and human whimsy will require a return to Jackman and regrouping for the second day along ME Rte. #27 because of the demolition of the bridge over the Moose River at Holeb. Otherwise, we could enjoy a scenic, instructive and exciting route through the heart of the Chain Lakes massif via the Beaudry and Gold Brook lumbering roads to the Stratton/Eustis area. If time allows, two optional stops near Parlin Pond, south of Jackman off Rte. U.S. 201, will be added to look at Ordovician (?) and Devonian (?) igneous rocks of the Jackman region.

Day #2 will traverse Maine Rte. #27 from north to south where we will see rocks at the roadside or on short walks into the forest. This transect is coincidently along the North Branch of the Dead River and will emphasize rocks of the southwestern part of the Chain Lakes massif and rocks of the Boil Mountain ophiolite ramped upon it. The selection of stops on day #2 will be flexible combining the influence of time, interest, and weather. The timing of the trip will be planned to allow participants adequate time to reach Farmington for the annual banquet and formal part of the meeting.



DAY #1  
Friday 10/13/89

### TRIP A-4 ROAD LOG

#### Mileage

0.0      ASSEMBLY POINT      9:00 AM

"Attean View" scenic rest stop located 6.0 miles south of the Moose River bridge between Jackman and Moose River. The overlook is about 85 miles from Farmington. No camping is allowed here.

The overlook is on rocks of the Attean batholith (ca. 445 Ma), mostly granite and granodiorite, situated on the west side of Owls Head. A panorama of the Boundary Mountains and the upper Moose River basin is seen from here. The Boundary Mountains are underlain either by metagreywacke and greenstone of the Silurian Frontenac assemblage intruded by diorite sills (probably also of Silurian age); or Helekian granofels diamictite of the Chain Lakes massif. The lowlands are underlain mostly by rocks of the Attean batholith or the Lower Devonian Seboomook Formation which is mostly composed of cyclic turbidite.

Depart at 9:00 - - North on US #201.

6.0      Moose River bridge.

9.0      Intersection of Holeb Road (unmarked) in Dennistown - - turn left toward the west.

15.4      N. Branch, Wood Stream.

16.6      Wood Stream    STOP #1 - - Rocks of the Attean batholith near Smith Pond.

21.1      Mud Pond Road Jct.    STOP #2 - - Rocks of the Chain Lakes massif; Burnt Jacket Mtn. type.

22.3      Road Jct. (Turner Ponds) - - Keep left.

23.0      McKenney Pond    STOP #3 - - Rocks of the Chain Lakes massif; McKenney Ponds type.

27.9      Road Jct. (Gulf Stream)    Turn right.

28.3      Pavements in road    STOP #4 - - Rocks of the Chain Lakes massif with segmented ribbon magnetite.

29.5      Fork in Gulf Stream Road - - Turn right.

30.1      Barren Ridge to South    STOP #5 - - Rocks of the Chain Lakes massif with amphibolite and other clasts.

30.6      Pavements along road    STOP #6 - - Rocks of the Chain Lakes massif with combination of "facies types".

Retrace route to Jackman

An announcement will be made at STOP #6 about two optional stops to be added between Jackman and points south. If these stops are added we will regroup at the starting point (Attean View rest stop) at a time that will allow for refreshment and fuel in Jackman. The optional road log is as follows:

0.0      Assembly Point    (Attean View)

(10.8)    Scott Paper Company Appleton Road (unmarked) near the south end of Parlin Pond - - turn right.

(15.0)    STOP #7 - - Lower Devonian (undated) garnet porphyry of the Moose River synclinorium.

(18.6)    STOP #8 - - Rocks of the Attean batholith - - "gray facies".

End of Day #1, Trip A-4 - - retrace route and proceed south. Participants continuing on the second day of the trip may wish to follow the leaders (who are familiar with time/mileage-saving short cuts) to Stratton.



DAY #2  
Saturday 10/14/89

## TRIP B-4 ROAD LOG

### Mileage

0.0      ASSEMBLY POINT      8:30 AM

Turnout southwest side of ME Rt. #27 about 500 feet northwest of access road to Natanis Point Campground.

Early arrivals and campers will enjoy the view to the southeast from Natanis Beach which displays a geologic transect from Lower Devonian (373 and 368 Ma) plutonic rocks intruding Heleikian Chain Lakes diamictite, over Cambrian ophiolite and related rocks passing upward into Siluro-Devonian rocks as young as Gedinian intruded by Lower Devonian gabbro and granitic rocks. This transect has the additional intrigue of being, in part, coincident with the MERQ/USGS Maine vibroseis/refraction geophysical line.

The leaders caution the participants that Rte. #27 presents us with special traffic dangers because of curves, narrow shoulders and traffic that mostly ignores speed limits. We have chosen turnouts with safety in mind, but still need your vigilance and care.

Stop locations can be identified on the (1) Chain-of-Ponds, (2) Jim Pond, and (3) Tim Mountain USGS 7.5-minute quadrangles; and (4) the Stratton 15-minute quadrangle. Quadrangle numbers are given in stop description.

1.0      STOP #1 - - Cliffs along east side of highway midway along Natanis Pond (1). Middle Devonian porphyritic granite of the Chain-of-Ponds pluton intruded by a Triassic(?) lamprophyre dike.

3.4      STOP #2 - - Cliffs on east side of highway opposite Bag Pond (1). Matrix-dominant massive granofels of the Chain Lakes diamictite about 1000 ft (300 m) from the contact with rocks of Stop 1.

4.8      STOP #3 - - Outcrops of east side of highway north of maintenance sheds near outlet of Lower Pond (1). Flecky gneiss of the Kibby Mountain facies of the Chain Lakes diamictite with abundant clasts.

6.3      STOP #4a - - Outcrop on east side of highway south of North Branch bridge (1). Massive, quartz clast variant of the Sarampus Falls facies which has been dated.

6.6      STOP #4b - - CLM (Sarampus Falls). Park Here.

Roadside and stream outcrops at Sarampus Falls Roadside Park (2). Typical Chain Lakes of diamictite facies, rheomorphic and partially layered showing anatectic leucosomes. The outcrop is northeast of a major zone of late brittle deformation.

8.8      STOP #5 - - Outcrops about 250 ft (75 m) east of highway, along indistinct fisherman's trail that leads to North Branch of the Dead River, opposite outlet of Viles Brook (2). Fault contact between Chain Lakes diamictite and ultramafic to epidiorite lens of the Boil Mountain ophiolite; includes, in part, a cumulate facies.

For STOP #6, see below (Optional)

- 10.1 STOP #7 - - Exploration pit (gold) about 100 ft (30 m) east of highway, at south intersection of a bypass road segment; about 1800 ft (550 m) north-northwest of Shadagee Falls (2). Cataclastic, altered trondhjemite of the ophiolitic Boil Mountain Complex.

Park between stops near Poison Pond

- 10.5 STOP #8- - Road cuts on both sides of highway; 600 ft (180 m) southeast of Poison Pond (2). Pillowed tholeiitic greenstone of the lower part of the Jim Pond Formation, where pillow facing direction is southeast.

Optional: Return to STOP #6 (Optional) \*Special navigational instructions and map to be furnished.

Cliff and ravine escarpment 1000 ft (300 m) west-southwest of the south end of Blanchard Pond (2). Same as Stop 5, with well-displayed fault relationships and chromite cumulate layers.

- 14.9 Jct. ME Rte. #27/CCC Road (unmarked)

NOTE #1: The following stops (9 through 12) will not be visited, but navigational directions are given for those who would like to pursue additional details on their own within the rocks of the Chain Lakes massif, Boil Mountain ophiolite, and Jim Pond Formation.

- 0.0 [14.9] Jct., CCC Road (unmarked) and ME Rte. #27 - - Turn left (Reset mileage to 0.0)

- 0.3 [15.2] Cross-roads - - Keep left.

- 0.7 [15.6] Jct. Keep left.

- 3.3 [18.2] Jct. Keep left. - - Jim Pond road (unmarked)

- 4.2 [19.1] Jct. at bridge ruins -- Keep right.

- 7.6 [22.5] STOP #9 - - Knoll (el. about 1290 ft) east of North Branch of Dead River, azimuth 320°, 1000 ft (300 m) from outlet of Viles Brook (2). Tectonite within the Sarampus Falls facies of the Chain Lakes diamictite formed by transposition and cataclasis.

(Turn and return)

- 8.1 [23.0] STOP #10 - - Cliff east of North Branch, Dead River, 1100 ft (330 m), azimuth 115° from outlet of Viles Brook. Epidiorite autobreccia of the Boil Mountain ophiolite.

- 9.7 [24.6] STOP #11 - - Outcrops east of North Branch, Dead River, 800 ft (240 m) north of Chase Pond along abandoned logging road on the southwest flank of Chase Pond Mountain (2). Keratophyric volcanoclastic rocks of the Jim Pond Formation, including vent facies breccia.

- 11.7 [26.6] STOP #12 - - Hill (el. 1291 ft) east of North Branch Dead River, 1600 ft (480 m), azimuth 100° from outlet of Shallow Pond along abandoned logging road



(2). Virginite associated with diapiric serpentinite along faults in uppermost greenstone of the Jim Pond Formation.

(Return to ME Rte. #27)

15.2 [30.1] ME Rte. #27/CCC Road - - Keep left.

NOTE #2: The following stop (13) and a diversion of about one mile on Eustis Ridge Road to the overlook and a pavement outcrop of the Hurricane Mountain mélange (same rocks as STOP #14) may be made depending on time available or if weather does not allow the traverse to STOP #6. Continue mileage from STOP #8.

15.8 STOP #13 - - A. Roadside outcrop at benchmark 1190, 0.1 mi (0.16 km) north of Eustis village, near north bypass road intersection (4). Pelitic member of the Dead River Formation. B. Dam at Eustis village (4). Calcareous graywacke member of the Dead River Formation.

16.9 STOP #14 - - Gravel pits 1800 ft (540 m), azimuth 115 m from Welhern Pond, about 0.5 mi (0.8 km) southwest of Eustis village (4). Scaly carbonaceous, sulfidic melange of the Hurricane Mountain Formation.

18.4 Jct., Me Rte. #27/Eustis Ridge Road

(End of Day #2, Trip B-4)

SIGNIFICANCE OF Al SILICATE IN STAUROLITE-  
GRADE ROCKS, CENTRAL MAINE

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## INTRODUCTION

In most medium-grade pelitic metamorphic terranes, staurolite first appears from reaction of garnet-chlorite-muscovite, or chloritoid. At slightly higher grades, the first Al silicate appears from reaction of staurolite-chlorite-muscovite. Generally, most of the staurolite zone includes rocks containing staurolite randomly interspersed with rocks containing staurolite and one of the Al silicates (kyanite, sillimanite, or andalusite), depending on pressure. There are several possible reasons for this apparently random distribution of Al silicate in staurolite-grade rocks. The Al silicate-bearing rocks could be the result of (1) reduced  $f(\text{H}_2\text{O})$ , (2) locally increased  $T$ , or (3) bulk compositional effects such as  $\text{Mg}/(\text{Mg} + \text{Fe})$ , here designated  $\text{Mg\#}$ , in percent.

Holdaway has been studying the differences between staurolite and staurolite-Al silicate schists in an effort to explain their distribution. In four different metamorphic terranes, the occurrences can be attributed purely to bulk compositional differences, and there is no necessity to call on alternate hypotheses ( $f(\text{H}_2\text{O})$  or  $T$  differences) to explain Al silicate occurrence. The results to date suggest that both  $T$  and  $f(\text{H}_2\text{O})$  behave in very predictable ways in most metamorphic rocks of medium grade.

The primary purpose of this field trip is to observe the distribution of staurolite and Al silicates in medium-grade schists which have undergone one or both of two metamorphic events (M2 and M3), and to illustrate the differences between compositional and thermal instability of staurolite. This field trip will include stops within the Augusta, Farmington, Anson, and Bingham quadrangles (Figs. 1, 2).

## METAMORPHIC HISTORY

The metamorphic history of central and west-central Maine has been summarized by Guidotti (1970, 1974), Holdaway et al. (1982, 1988), and Dickerson and Holdaway (1989). Only two of the four metamorphic episodes are of concern here, M2 and M3. These events are mainly regional in scale but owe their origin to thermal effects associated with plutons of the New Hampshire Magma Series.

M2 is defined as an early event (about 400 Ma) in which staurolite reacted thermally to produce andalusite-bearing assemblages, with sillimanite



appearing after staurolite had disappeared (Dickerson and Holdaway, 1989). In addition to staurolite and andalusite, cordierite occurred in some rocks which experienced this event. Dickerson and Holdaway suggest that P was about 2.7 kbar in regional M2 rocks of the Bingham area. Contact metamorphic variants are cordierite-rich (M2n, 2.35 kbar), and staurolite-bearing (M2s, 2.75 kbar) as shown in Figure 1. Almost all M2 rocks are slightly to fully retrograded by more recent events.

M3 is defined as an event which took place between 394 and 379 Ma in which staurolite reacted thermally to produce sillimanite-bearing assemblages, with no evidence of intervening andalusite (Holdaway et al., 1982, 1988). M3 is well-developed in Augusta and southern Farmington quadrangles as well as most other areas between Augusta, Rumford, Rangeley, and Madison (Fig. 2). Holdaway et al. (1988) estimate P of M3 to be about 3.1 kbar.

In Augusta quadrangle staurolite breaks down directly to sillimanite without intervening andalusite, but andalusite is widespread among staurolite-grade rocks. This andalusite has its origin in M2, but, as will be shown below, it remained stable during M3 and began reacting to sillimanite at the same grade at which staurolite began producing sillimanite.

### ASSEMBLAGE DESIGNATIONS

In order to illustrate the differences between occurrences of Al silicate resulting from thermal instability of staurolite and those resulting from compositional instability of staurolite, an alpha-numeric scheme is utilized which takes into account both grade and composition (Figs. 4, 6, 7). However, this scheme does not designate which Al silicate, andalusite or sillimanite is present; the specific Al silicates are designated in the caption.

Grade is designated by a number: 4 is staurolite zone, 5 is transitional to Al silicate zone, 6 is Al silicate zone. Rocks of grade 4 may be staurolite-biotite-garnet-muscovite, or that assemblage plus an Al silicate. Rocks of grade 5 commonly contain staurolite and must contain Al silicate; they must be spatially located between grades 4 and 6. Rocks of grade 6 cannot contain staurolite but do contain andalusite and/or sillimanite plus muscovite. Quartz is present in all assemblages and available for all reactions. Other accessory phases are also present but excluded from this discussion.

The presence of Al silicate vs. staurolite is indicated by A, B, or C. Thus 4A is a staurolite-grade rock without Al silicate, 4B contains staurolite and Al silicate, and 4C contains Al silicate and no staurolite, but all three assemblages are spatially distributed among each other at the same metamorphic grade. All possible medium-grade assemblages are listed below; those of equivalent grade are given on the same line.

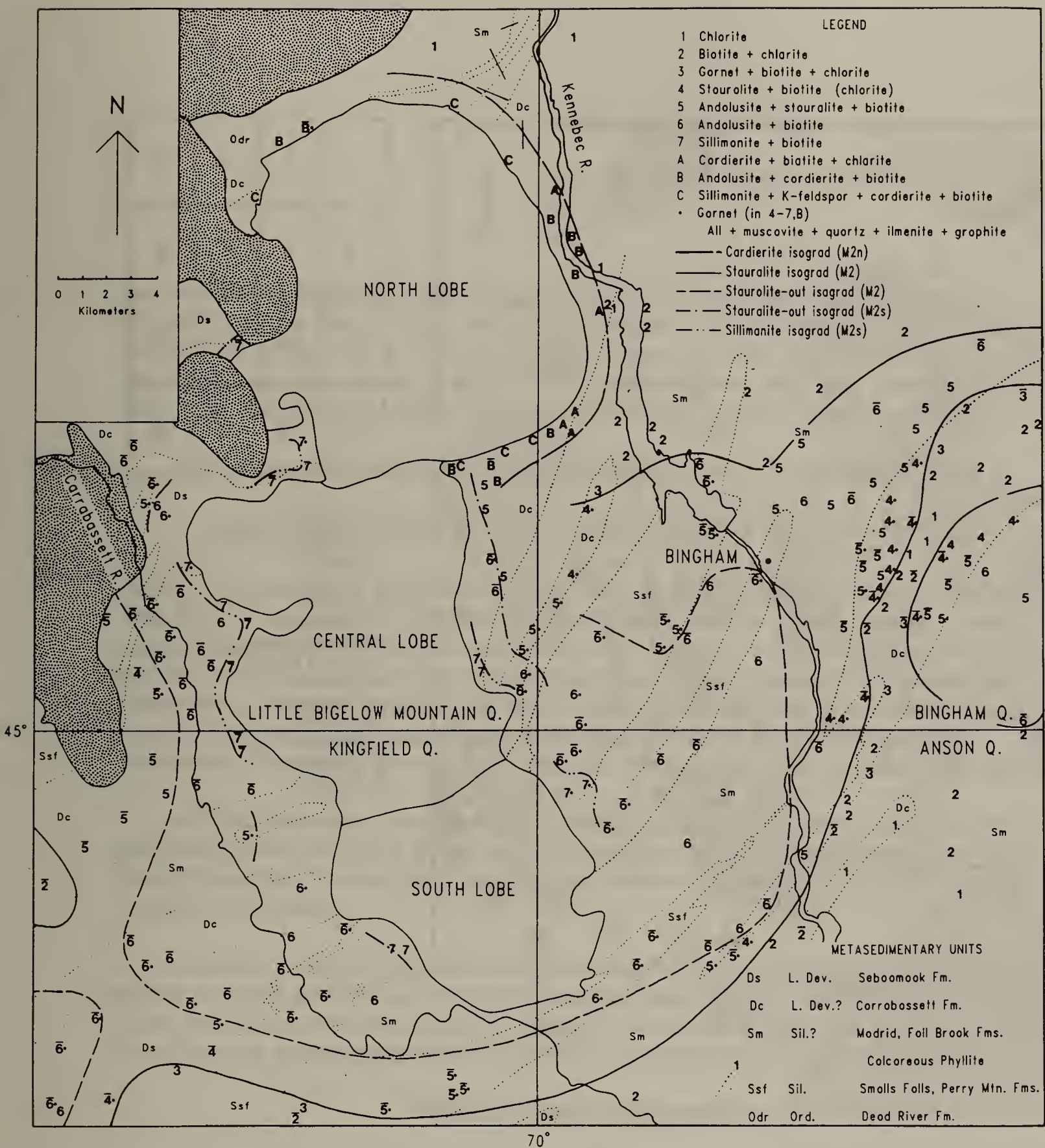


Fig. 1. Mineral assemblage and isograd map for M2 in the region around Bingham. Letters and numbers represent mineral assemblages as given in the legend. Within the staurolite zone assemblages 4, 5, and 6, correspond to 4A, 4B, and 4C in the text. Assemblage 5 immediately below the staurolite-out isograd corresponds to 5B in text, transitional to the Al silicate zone. Overbar indicates highest-grade minerals more than 90% retrograded. Acidic plutonic rocks - white, gabbroic - stipple. Reprinted with permission from American Journal of Science (Dickerson and Holdaway, 1989).



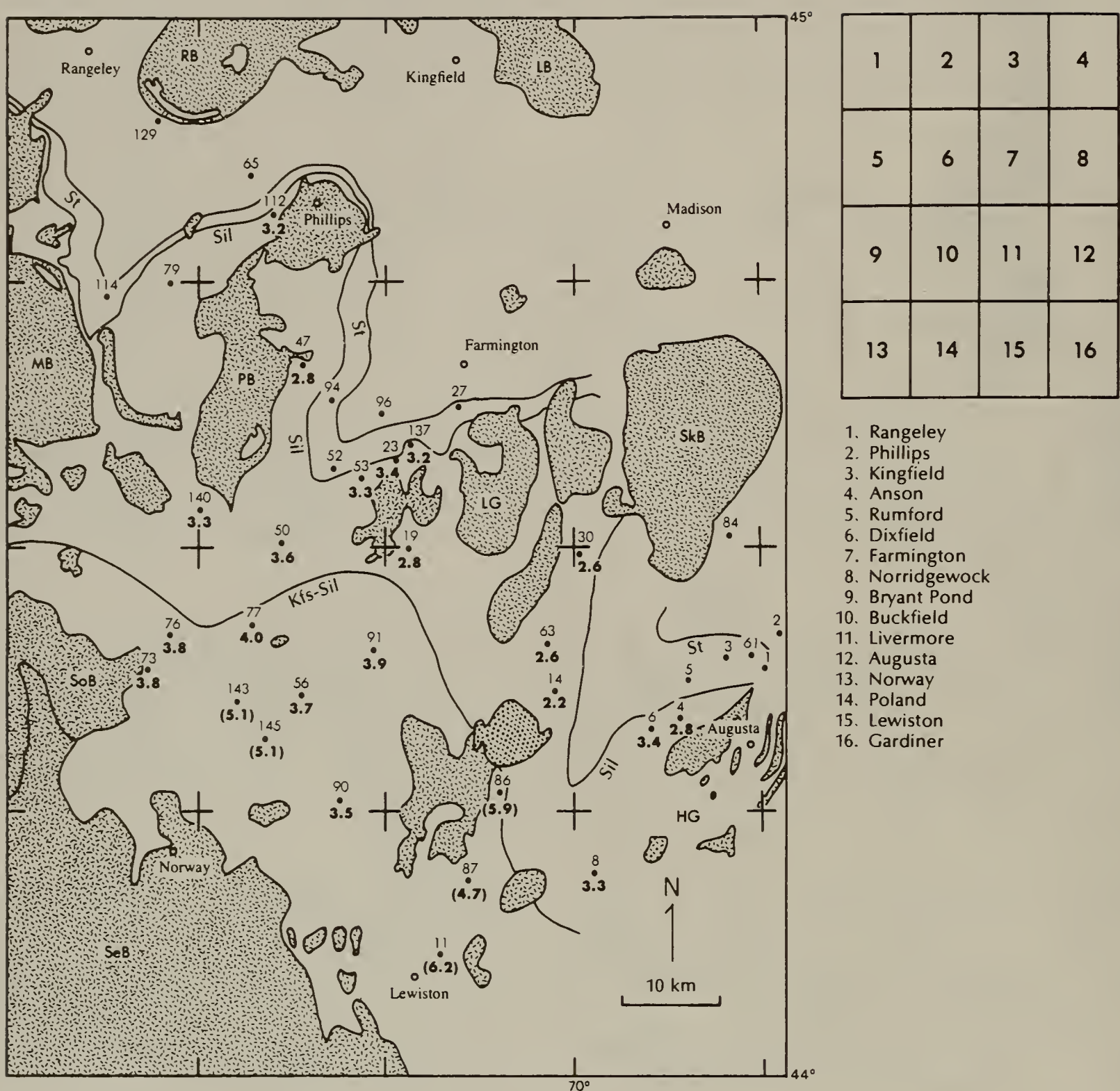


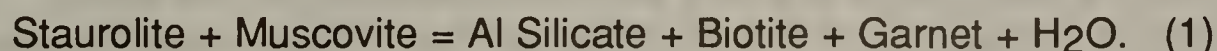
Fig. 2. Isograd map for M3 and M5 events in west-central Maine. Isograds shown are M3 staurolite (St), M3 sillimanite (Sil), and M5 K feldspar-sillimanite (Kfs-Sil). Whole numbers refer to analyzed specimens; bold decimal numbers refer to P determinations using the MABS geobarometer. Pressures in parentheses may represent disequilibrium. Patterns indicate plutonic rocks. Reprinted with permission from American Mineralogist (Holdaway et al., 1988).

## Composition

Grade	4A, 4B, 4C
	5B, 5C
	6C

Assemblages 4C and 5C are uncommon because pelitic rocks of staurolite grade rarely occur without staurolite. It is important to emphasize two aspects of grade 5: (1) such rocks must always contain Al silicate, and (2) they represent the transitional grade at which staurolite is thermally reacting away, and must therefore be spatially located between rocks of grades 4 and 6.

The reaction whereby staurolite breaks down in most pelitic rocks is



Ideally, the presence of garnet should help to identify rocks of assemblages 5B and 5C. However, the stabilizing effect of Mn and Ca allows garnet to exist in assemblages 4B and 4C as well as 5B, 5C, and 6C. Note that assemblages 4C, 5C, and 6C are identical in terms of the minerals present, as are 4B and 5B. However, study of Figures 4, 6, and 7 shows that they can be distinguished by map distribution, and we will see later that there are significant mineral chemical differences between 4B and 5B, and presumably between 4C and 5C.

The distinction between the compositional and thermal instability of staurolite is illustrated in the concluding section with a TX diagram (Fig. 8), which illustrates the similarity of mineral assemblages resulting from the two types of instability.

Each of the three areas visited in the field trip offers unique contributions to the concept of thermal vs. compositional occurrence of Al silicate. All pelitic rocks in the region formed under reducing conditions; graphite and hematite-free ilmenite are present in all specimens. Details of these three areas follow.

## AUGUSTA QUADRANGLE

In the Augusta quadrangle, Novak and Holdaway (1981), and Holdaway et al. (1988) have suggested that both M2 and M3 affected the area, based primarily on abundance of low-variance assemblages, some polymetamorphic textures, and the coexistence of minerals such as staurolite and cordierite which are incompatible elsewhere. However, it is not possible to distinguish two sets of isograds, implying that the two events were temporally close, perhaps representing a simple P increase during metamorphism.

Despite the fact that staurolite reacts directly to sillimanite, andalusite is very common and cordierite is not uncommon in normal pelitic rocks. Most, if not all, of the minor chlorite found in rocks of grades 4, 5, and 6 is retrograde.



This conclusion is suggested by chlorite textures and confirmed by the widespread occurrence of Al silicate-biotite and cordierite, which are both compositional alternatives to chlorite. The reaction



destroyed chlorite very early in grade 4 and left andalusite-biotite as the widespread alternative to staurolite-chlorite. The origin of chlorite in medium-grade rocks of M3 is discussed in detail by Holdaway et al. (1988).

In the Augusta area there are several occurrences of the assemblage staurolite-andalusite (or sillimanite)-biotite-cordierite-muscovite-garnet-chlorite, two of which will be examined on the trip. The garnet is minor and is stabilized by Mn + Ca, and the chlorite is retrograde, as discussed above. However, the presence of staurolite-andalusite-cordierite-biotite-muscovite is an enigma. Because Al silicate-biotite is a very common assemblage in all Maine metamorphic rocks, even in the staurolite zone, we believe staurolite and cordierite are not compatible in rocks dominated by AFM components. Novak and Holdaway (1981) suggested that Augusta area rocks experienced early M2 with common AFM assemblages being staurolite-andalusite-biotite-muscovite and andalusite-biotite-cordierite-muscovite, followed immediately by M3 with increasing P producing staurolite-biotite-garnet-muscovite and partially to totally destroying andalusite and cordierite (Fig. 3). In some specimens, this process left the polymetamorphic assemblage present at West Sidney (Stop 2).

If the rocks contained significant amounts of non-AFM and non-saturated components, then there is the possibility that staurolite-andalusite-cordierite-biotite-muscovite might locally have been stable. Such a case probably occurs at East Winthrop (Stop 3) where Zn stabilizes staurolite and Li stabilizes both staurolite and cordierite relative to andalusite-biotite (Dutrow et al., 1986).

Figure 4 shows the distribution of andalusite in staurolite-grade rocks of the Augusta area. Throughout M3, staurolite, biotite, and garnet of assemblage 4A have consistently lower Mg# than those minerals of assemblage 4B. In addition, the garnets of assemblage 4B contain more Mn + Ca than those of 4A (Fig. 5). This suggests the following: (1) andalusite was an equilibrium phase in these M3 rocks despite its probable formation during M2; (2) the appearance of andalusite in M3 is related to local variation in bulk composition; (3) the limiting Mg#, beyond which andalusite-biotite must form instead of more Mg-rich staurolite, is more or less constant over the T range of the staurolite zone. In the transitional rocks of assemblage 5B, staurolite of all compositions is breaking down; therefore there is little compositional control on the first appearance of sillimanite. Biotite, staurolite, and garnet with a wide range of compositions occur in grade 5B (Fig. 5). Comparable ranges of biotite and garnet compositions occur in the staurolite-free rocks of grade 6C. In M3 the sillimanite isograd coincides with the beginning of the thermal breakdown of staurolite (5B), and this combined with garnet-biotite geothermometry provides an accurate way to estimate pressure (Holdaway et al., 1988).

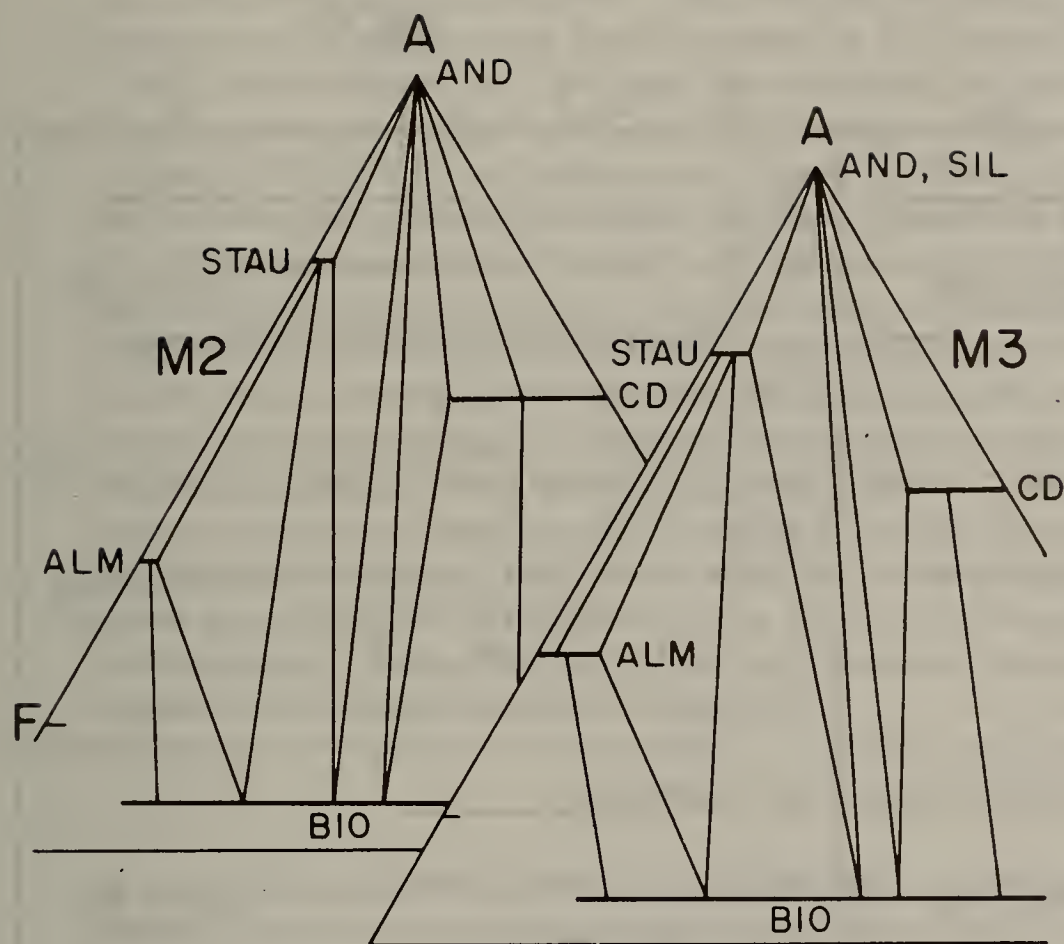


Fig. 3. Schematic AFM diagrams for medium-grade M2 and M3. As pressure decreases, the Al silicate-biotite tie lines rotate clockwise, decreasing the compositional field of staurolite, and increasing the compositional field of cordierite. M2n (Fig. 1) corresponds to staurolite-unstable conditions with a still-greater clockwise rotation of tie lines than that shown for M2.

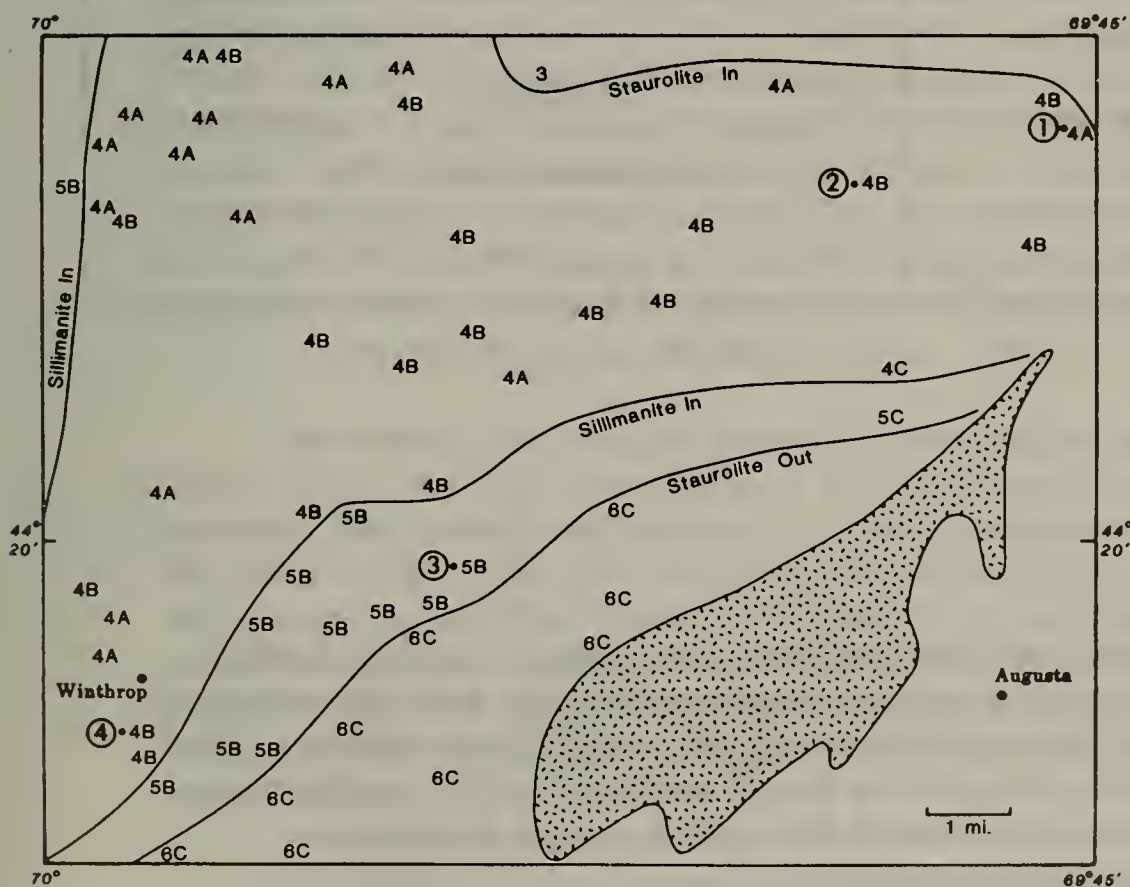


Fig. 4. Mineral assemblage and isograd map for M3 of the Augusta area. Mineral assemblages are defined in the section entitled ASSEMBLAGE DESIGNATIONS. Also, 1 - chlorite z., 2 - biotite z., 3 - garnet z. Circled numbers refer to field trip stops. Pattern shows granitic rocks. All 5B and 6C assemblages contain sillimanite, but a few also contain relict andalusite.



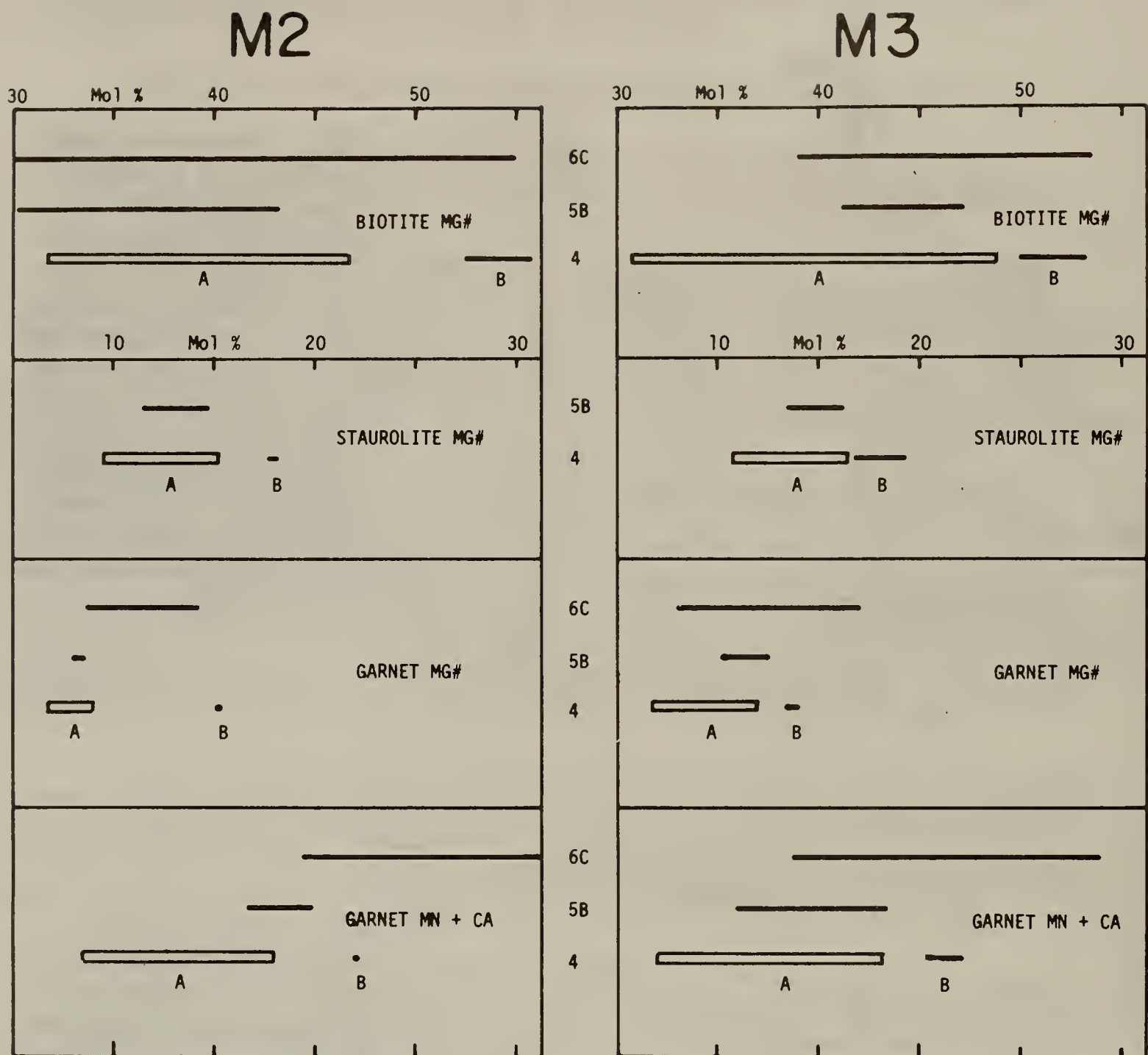


Fig. 5. Diagrams to illustrate ranges of mineral compositions for M2 and M3. Assemblages are designated in the margin between the M2 and M3 rectangles. Assemblage 4A is shown by an open bar; 4B, 5B, and 6C are shown by solid lines. Note that 4A and 4B compositions do not overlap, but 5B overlaps 4A. Garnets of 6C are commonly enriched in Mn + Ca because almandine component is reacting away at that grade.

In the Farmington area (discussed below) most M3 was not preceded by M2, and andalusite is absent. The P of M3 in the Augusta and Farmington areas was identical within limits of error at 3.1 kbar (Holdaway et al., 1988). These observations and the fact that cordierite is locally developed in the Augusta area support Novak and Holdaway's (1981) conclusion that M2 preceded M3 in the Augusta area. However, the strong compositional control of andalusite occurrence discussed above implies that the andalusite which is now preserved remained stable with coexisting biotite and staurolite during M3. We attribute this preservation of andalusite to mosaic equilibrium during M3. After M2, local differences in bulk composition existed near larger andalusite grains. As P increased to M3 conditions andalusite and biotite reacted to make more Mg-rich staurolite. Where andalusite was minor or fluids abundant, andalusite disappeared leaving assemblage 4A. Where there were larger andalusite grains, they tended to buffer the staurolite and biotite at their most Mg-rich compositions, and leave local domains with higher bulk Mg#. The same argument can be applied to the more local occurrences of cordierite. The destruction of cordierite and growth of staurolite stable with andalusite and biotite were arrested at some localities.

### FARMINGTON QUADRANGLE

In the southern part of the Farmington quadrangle, andalusite is virtually absent; the only specimen containing andalusite is in the northern part of the staurolite zone, 4 km south-southeast of Farmington (Fig. 6). Most of southern Farmington quadrangle has been affected by M3, but highly retrograded M2 rocks occur along the western edge and are truncated by M3 isograds in the southwestern part of the quadrangle (Dutrow, 1985). The M2 effects increase in grade westward and andalusite appears in the adjacent Dixfield quadrangle (Henry, 1974). Cordierite is completely absent from the Farmington quadrangle. Thus the assemblages in M3 here are almost exclusively 4A, 5B, and 6C, with sillimanite as the only Al silicate. The first appearance of sillimanite marks a distinct isograd (Fig. 6), and there is no random distribution of 4A and 4B assemblages as occurs in Augusta quadrangle and in M2 rocks to the north in the Kingfield-Bingham region. The first sillimanite appears at about 560° in both quadrangles (Holdaway et al., 1988).

We attribute the absence of 4B assemblages to the absence of M2 effects prior to M3. The bulk compositions of Farmington (and Augusta) area pelites are such that a single M3 event (3.1 kbar) produces only assemblage 4A in the staurolite zone. This restricted range of mineralogy may be changed in one of three ways to produce 4A and 4B together: (1) by absence of graphite, more hematite-rich ilmenite, and therefore less Fe in the silicates; (2) by presence of sulfides to also tie up the Fe in non-silicates or (3) by an earlier lower P metamorphism to produce andalusite-rich domains which remain richer in Mg as discussed above.

In addition to the regular grade sequence, the Farmington quadrangle exemplifies well the control of isograds by adjacent plutonic bodies (Fig. 6). We



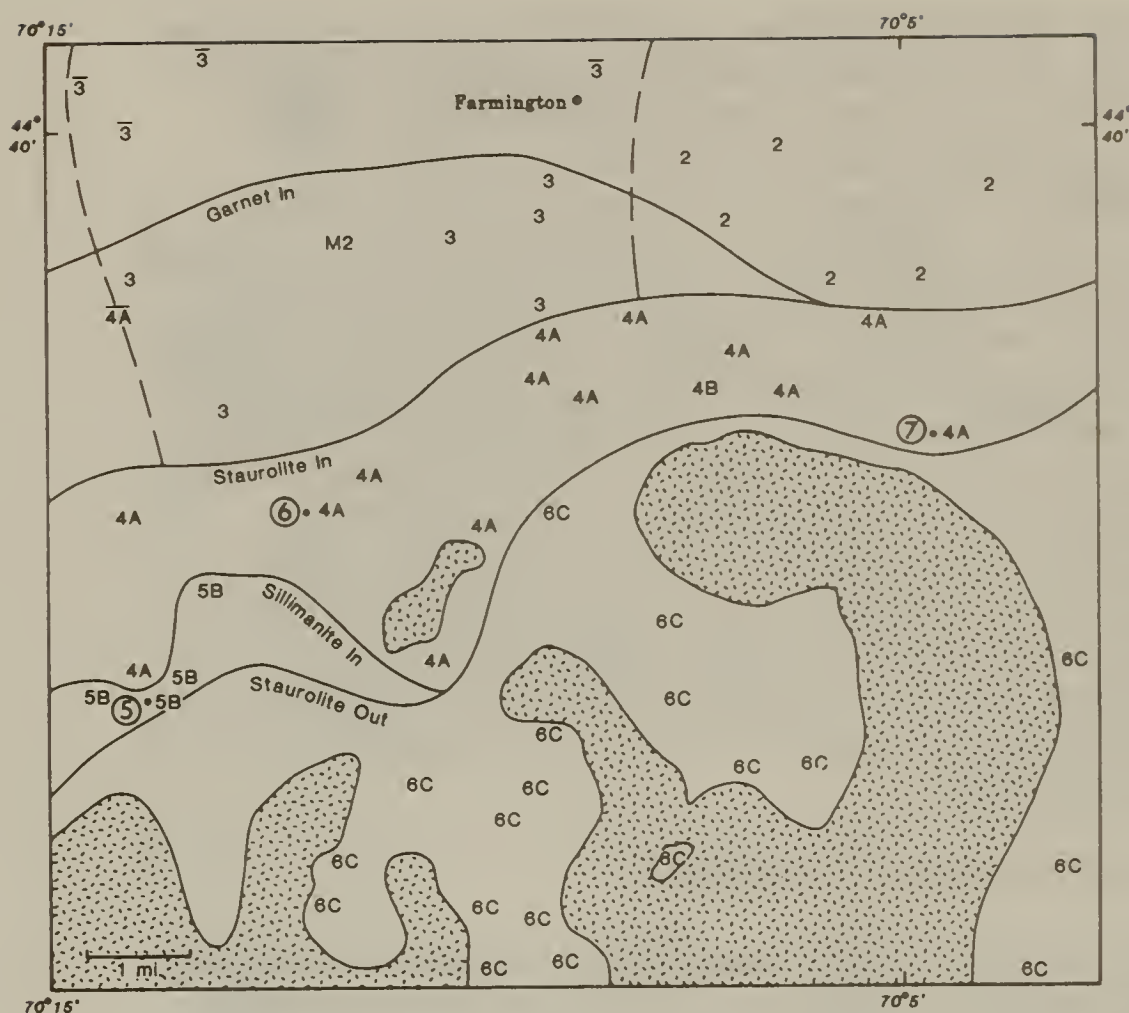


Fig. 6. Mineral assemblages and isograd map for M3 of the Farmington area. Dashed lines are M2 isograds. All 5B and 6C assemblages contain sillimanite; the only andalusite occurs at the one 4B locality. See caption of Figure 4 for additional details.

believe the plutonic bodies came in over a short time interval and P increased slightly southward (Holdaway et al., 1988).

East of Farmington and north of Augusta, in the northeastern part of Norridgewock quadrangle, the contact metamorphism near the Skowhegan batholith shows indications of P lower than that for M3 near Farmington. In this eastern area, andalusite and cordierite are both more prevalent than staurolite. There is no evidence for polymetamorphism in these rocks, and the isograds are continuous with M3 isograds to the west and south. The interpretation of this area was left open by Holdaway et al. (1988) as shown on Figure 2, but it now appears that the area is best interpreted as a northeastward extension of M3 at anomalously low P. The metamorphism might have begun during M2 time, but continued into M3 time without a P increase. This area presents a challenge to our method of identifying events. Strictly speaking, it should not be interpreted as M3 because it has P more typical of M2; however, the lack of truncation of isograds and possible synchrony of time with M3 supports the determination as M3.

#### ANSON AND BINGHAM QUADRANGLES

The M2 rocks of Anson, Bingham, Kingfield, and Little Bigelow Mountain quadrangles formed at lower P (about 2.7 kbar) than the younger M3 rocks to the south, evidenced by the presence of andalusite as a product of staurolite breakdown (Reaction 1) instead of sillimanite (Dickerson and Holdaway, 1989).

Andalusite is also very widespread among staurolite-grade rocks, assemblages 4A, 4B, and 4C (Figs. 1, 7) all being present. (Note that assemblages 4, 5, and 6 in Figure 1 all intermixed in the staurolite zone represent assemblages 4A, 4B, and 4C respectively in Figure 7).

Cordierite is totally absent from the regionally developed M2 and from the slightly younger contact metamorphic rocks (M2s) around the central and south lobes of the Lexington batholith; however cordierite is widespread in contact metamorphic rocks around the (earlier?) northern lobe of the batholith (M2n, 2.35 kbar), and staurolite is absent from these northern rocks (Fig. 1). Throughout the region of the Lexington batholith, staurolite and cordierite are mutually exclusive. The alternative assemblages staurolite-andalusite-biotite and cordierite-andalusite-biotite are very common. These observations provide strong evidence that staurolite-cordierite-biotite-muscovite is not a stable assemblage except possibly in the presence of significant amounts of non-AFM components such as Zn and Li.

As with M3, the M2 staurolite, biotite, and garnet of assemblage 4B are more Mg-rich than those minerals in 4A, and in addition the garnets contain higher Mn + Ca (Fig. 5). The 4B assemblage corresponds to bulk compositions with higher Mg# or higher  $Al/(Al + Mg + Fe)$  as shown in Figure 3. Some garnet persists in 4B as a result of its increased Mn + Ca which stabilizes the andalusite-biotite-garnet assemblage of Reaction 1 to lower T along with staurolite. The 5B compositions overlap 4A and 4B as is the case for M3 (Fig. 5). The most Mg-rich staurolite, biotite and garnet of assemblage 4A in M2 are slightly more Fe-rich than analogous minerals of that assemblage in M3, suggesting that staurolite in pelitic rocks is stabilized to more Mg-rich compositions by increased P. For this reason andalusite is much more prevalent in these M2 rocks than in the Augusta area.

Stop 9 in Bingham quadrangle represents the only known locality in M2 or M3 where chlorite occurs with staurolite in significant amounts and has a foliated texture suggesting that it is a prograde mineral. The rocks here also contain andalusite. This locality represents the lower part of the staurolite zone, 300 m from rocks transitional to the garnet zone. Staurolite is reacting with chlorite to produce andalusite-biotite (Reaction 2). The fact that this locality represents the lower part of the staurolite zone and the beginning of the disappearance of staurolite-chlorite reinforces our contention that throughout much of the staurolite zone Al silicate and biotite represent the compositional alternative to more Mg-rich staurolite, and chlorite in most such rocks is retrograde.

## CONCLUSIONS

The combination of chemical and field data from M2 and M3 in central and west-central Maine provides for a number of useful petrologic conclusions. These may be applied to the interpretation of metamorphic history in Maine and elsewhere.



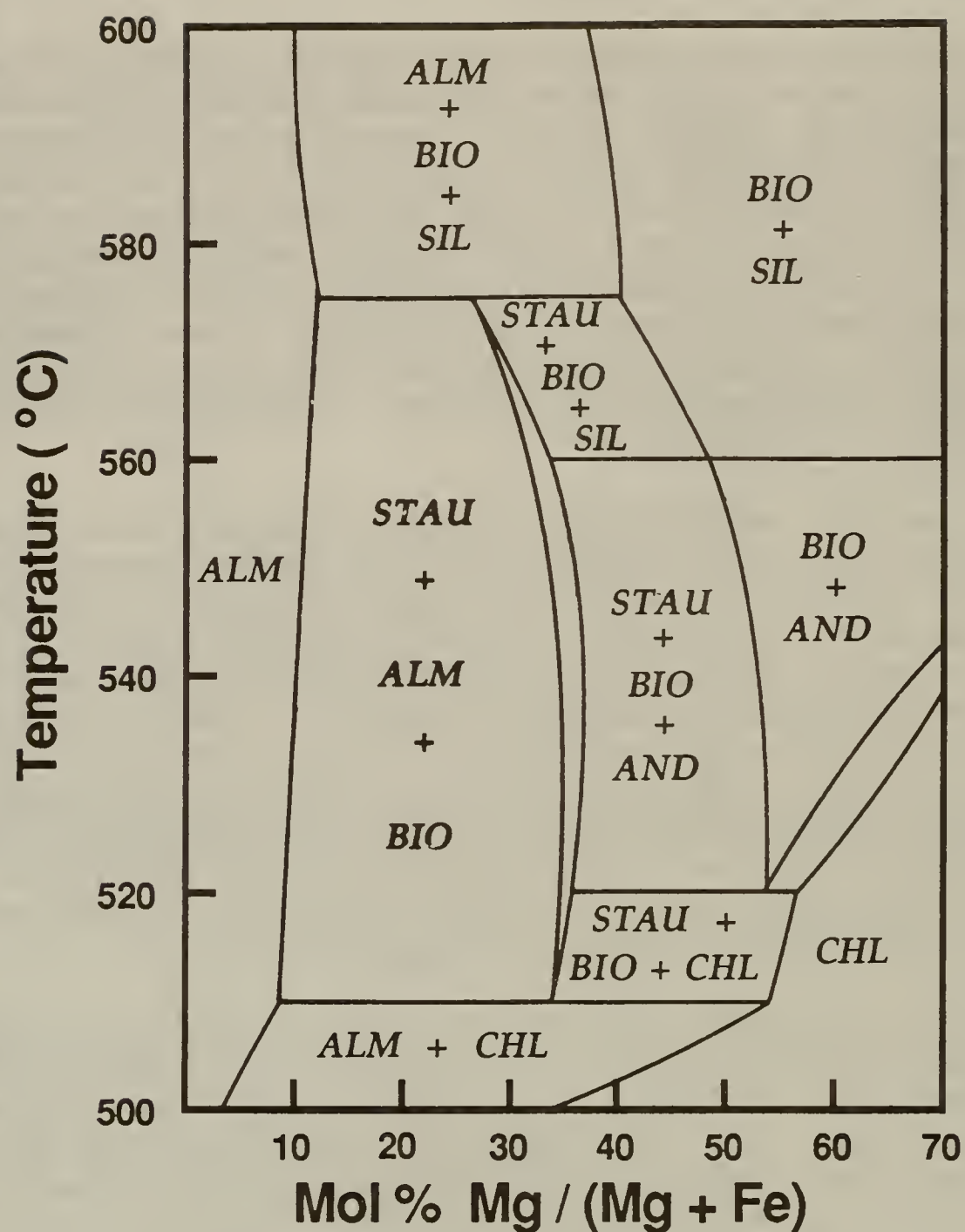


Fig. 8. Schematic TX diagram for the almandine-chlorite join at 3.1 kbar based on compositions and geothermometry from M3 in Maine. At this pressure, biotite-sillimanite is the thermal alternative to staurolite and biotite-andalusite is the compositional alternative. Garnet is present with both assemblages, stabilized by Mn + Ca in the latter. Chlorite is the compositional alternative to staurolite near the beginning of the staurolite zone. Narrow sliver at 35% = staurolite-biotite; narrow sliver at 60% = biotite-andalusite-chlorite.

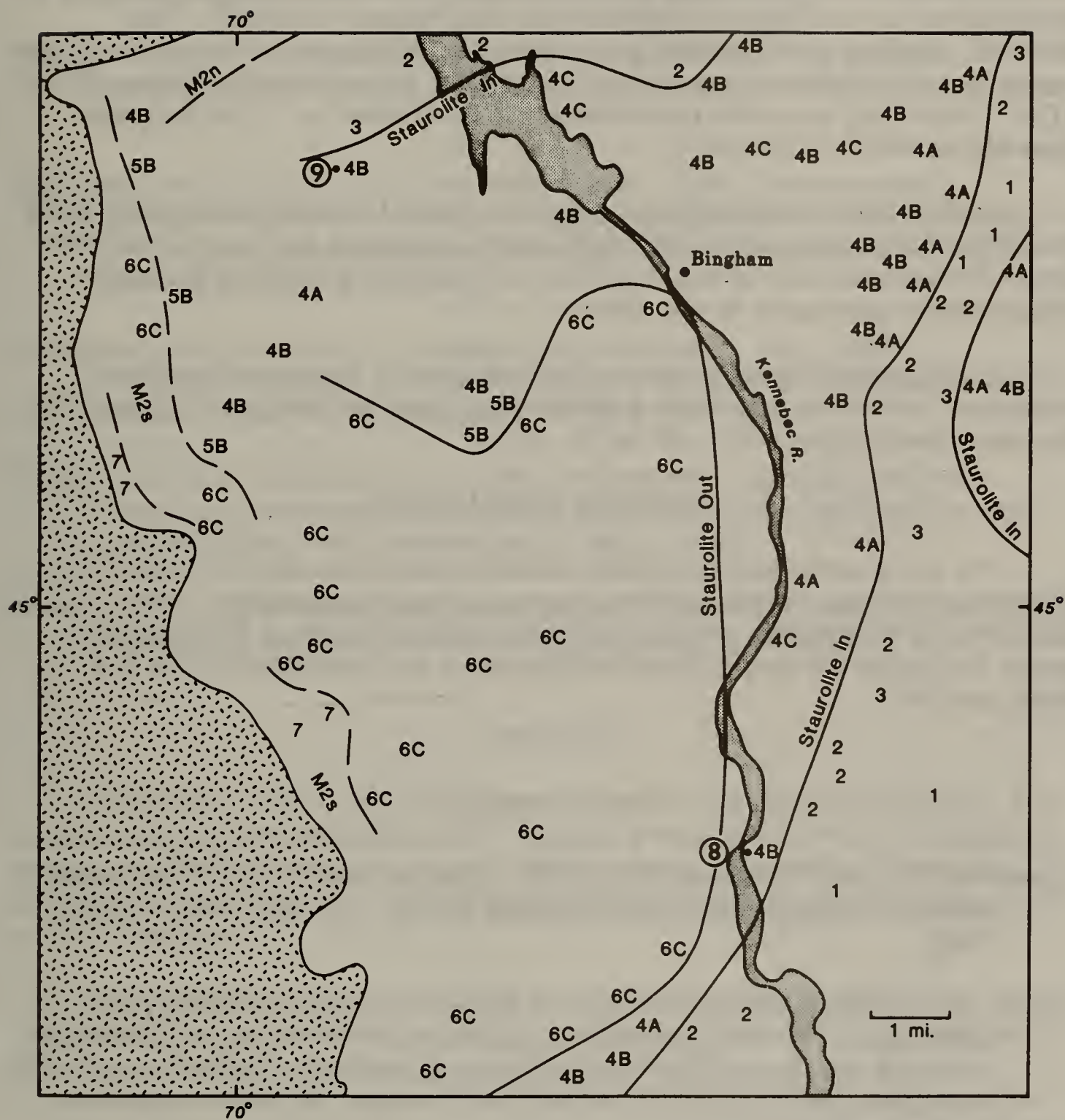


Fig. 7. Mineral assemblage and isograd map for M2 of the Anson and Bingham quadrangles. Mineral assemblages shown correspond to M2 and M2s, which occurred at similar P. No isograd is shown for the disappearance of assemblage 4A, but localities immediately below the staurolite-out isograd are transitional to the Al silicate zone, and shown as 5B. Sillimanite does not appear until assemblage 7, the sillimanite-bearing equivalent of 6C. See caption of Figure 4 for additional information.



1. Al silicate-bearing assemblages may be produced by either compositional or thermal instability of staurolite. These two possibilities may be distinguished on the basis of field relations and mineral chemistry. The chemical relations are illustrated in a schematic TX diagram in Figure 8. When garnet forms in a field not labelled garnet, it must be stabilized by increased Mn + Ca. The limiting staurolite composition is little affected by T, but becomes more Mg-rich with increasing P.

2. Chlorite is an uncommon prograde mineral in staurolite-bearing rocks throughout the region and is only stable in the low-grade part of the staurolite zone. Throughout most of the staurolite zone Al silicate-biotite is the stable compositional alternative to staurolite.

3. Staurolite and cordierite do not coexist as a stable assemblage in micaceous schists except where stabilized by significant amounts of non-AFM, non-saturated components such as Zn and Li.

#### ACKNOWLEDGEMENTS

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## ITINERARY

Assembly Point: Rest Area on southbound I-95 near West Sidney, 3.8 miles south of Lyons Road (Exit 32). Assembly time is 8:00 A.M. Topographic maps: Augusta, Farmington, Anson, and Bingham. We will pass through Livermore quadrangle and the northwest corner of Norridgewock quadrangle without stopping.

In the guide, only primary AFM phases are listed in assemblages. All specimens contain muscovite-quartz-ilmenite-graphite-tourmaline, and some contain plagioclase. Number in parentheses designates grade and assemblage as defined under ASSEMBLAGE DESIGNATIONS in text. Also: 1 = chlorite zone; 2 = biotite zone; 3 = garnet zone; 7 = sillimanite-muscovite zone in M2s.

## Mileage

- 0.0 STOP 1. Roadcut immediately south of Rest Area on west side of I-95. CAREFUL! Pelitic rocks of the Waterville Formation at the staurolite isograd (?). At north end are garnet-biotite-chlorite schists (3). Green color indicates chlorite, interpreted to be primary. At 300 m south are staurolite-biotite-garnet schists (4A). The green color disappears, but garnet content does not decrease noticeably. The Mg# of staurolite is 14.9, that of biotite is 47.8, showing that the mineral



chemistry is within the range of 4A for M3 (Fig. 5). Return to cars and drive south on I-95.

- 3.9 First roadcuts of Hallowell Granite, a biotite muscovite quartz monzonite which may have acted as a heat source for M3.
- 4.4 Turn right at Exit 31 and proceed north on ME-8-11-27.
- 5.0 Sillimanite-andalusite biotite-garnet schist (6C) at the sillimanite isograd.
- 5.6 Andalusite-biotite-garnet schist (5C).
- 8.1 STOP 2. Take right fork on ME-23 and park on right beyond low roadcut. West Sidney locality, staurolite-grade rocks of the Waterville Formation about 2 km south of the staurolite isograd. Roadcut at intersection is mainly biotite-quartz-rich rocks. At 60 m south on east side are exposures of staurolite-andalusite-biotite-cordierite-garnet schist (4B) and subsets of this assemblage, with minor retrograde chlorite. A specimen of the maximum 4B assemblage collected from the middle of the outcrop has staurolite with Mg# 19.2 and biotite with Mg# 52.8 (Fig. 5). Staurolite of the 4B assemblage does not contain high Zn or Li. At 500 m south are roadcuts of staurolite-biotite-garnet schist (4A) and staurolite-andalusite-biotite-garnet schist (4B).  
  
There is an interesting evolution of thought regarding the large number of phases in the maximum 4B assemblage. Osberg (1971) suggested that the whole was an equilibrium assemblage. Novak and Holdaway (1981) suggested two equilibrium assemblages: M2 - andalusite-biotite-cordierite; M3 - staurolite-biotite-chlorite. We now think there were three phases of metamorphism: M2 - andalusite-biotite-cordierite; M3 - staurolite-andalusite-biotite; retrograde stage - chlorite. By any interpretation, the trace amount of garnet was stabilized by Mn. Turn around and reverse direction.
- 8.7 Turn right on Summer Haven Rd.
- 9.0 Take left fork on Old Belgrade Rd.
- 12.1 Proceed south on ME-135.
- 12.7 Take right fork on Puddle Duck Rd.
- 14.4 Turn right (west) on US-202, ME-11-100 at Manchester.
- 16.3 Turn right on side road.
- 16.4 Turn right on Case rd.

- 16.9 STOP 3. Park on hill at right beyond roadcut. East Winthrop locality, semipelitic rocks of the Waterville Formation south of the sillimanite isograd. Several of the rocks contain sillimanite, biotite, and garnet (5C). A single more pelitic specimen was collected from about 15 m south of the north end of the roadcut, a staurolite-sillimanite-biotite-garnet schist (5B). John Ferry has collected a similar specimen which also contains cordierite. Both the staurolite and the cordierite contain about 1% Li<sub>2</sub>O, and the staurolite also contains 1% ZnO (Dutrow et al., 1986). The staurolite has Mg# 12.9, and the biotite has Mg# 48.3, but the high Li has affected the Fe-Mg partitioning (Holdaway et al., 1988). The staurolite and cordierite are minor phases, and may be stabilized together with micas by the Li and Zn. Turn around in next driveway and return to highway.
- 17.4 Turn right on side road.
- 17.8 Rejoin US-202, ME-11-100, turn right, and proceed west.
- 18.6 Staurolite-sillimanite-andalusite-biotite-garnet schist (5B).
- 20.8 Biotite schist
- 21.9 Turn right (north) on ME-133 at Winthrop.
- 22.0 STOP 4. Park on right. Low roadcut on left of staurolite-zone rocks of the Sangerville Formation. At the southwest end, staurolite-biotite-garnet schist (4A) is most common, and at the northeast end staurolite-andalusite-biotite-garnet schist (4B) is more common. Andalusite occurs as a few large sievy crystals. This locality is in the reentrant between the sillimanite isograd of the Hallowell lobe of M3 and the Livermore Falls lobe of M3 (Figs. 2, 4). A 4A specimen contains staurolite with Mg# 15.0 and biotite with Mg# 47.1. Continue north and west on ME-133.
- 23.2-23.5 Staurolite-andalusite-biotite-garnet schist (4B) on left. The only andalusite seen in M3 of the Livermore Falls lobe is along the eastern edge, suggesting that earlier M2 decreased in grade west from here.
- 24.5 Enter Livermore quadrangle on ME-133.
- 25.5 Staurolite-sillimanite-andalusite-biotite-garnet schist (5B).
- 28.0 Sillimanite-biotite-garnet schist (6C).
- 28.4 Wayne, stay on ME-133. Several poorly exposed two-mica granites provided heat for the Livermore Falls lobe of metamorphism (Fig. 2). A gabbroic pluton of Carboniferous age is poorly exposed south of Wayne.



- 30.4 Biotite quartzites exposed on left.
- 38.6 ME-17 joins ME-133 from the right, continue northwest through Livermore Falls on ME-17-133.
- 39.4 ME-133 turns right, continue northwest on ME-17 through Chisholm and Jay.
- 41.8 Enter Farmington quadrangle on ME-17-4.
- 45.6 ME-17 turns left, continue north on ME-4.
- 48.1 Turn left (west) on US-2.
- 48.5 Turn right here for lunch materials in Wilton. (Foodland 0.5 miles on left).
- 49.2 LUNCH STOP. Turn left into Rest Area. After lunch reverse direction, east on US-2.
- 50.1 STOP 5. Park on right adjacent to large roadcuts. Lower sillimanite zone rocks of the Sangerville Formation. Most of the pelites are staurolite-sillimanite-biotite-garnet schists (5B) of M3. The schists locally contain coarse muscovite pseudomorphs of staurolite (spangles) described by Guidotti (1968). In such specimens, most of the staurolite occurs as remnants within the muscovite pseudomorphs. Spangles are best seen near the middle of the exposure on the left side. At a locality east of here, the spangles occur with very little sillimanite in the rock, implying that K-bearing fluids from the nearby pluton affected the rock. A specimen from near the locality of Stop 5 has staurolite with Mg# 13.5 and biotite with Mg# 41.2, typical of 5B specimens (Fig. 5). Stop 5 is about 2 km north of a two-mica granitoid (Fig. 6), and the M3 isograds trend east-northeast in this area paralleling the contacts of granitic plutons. About 5 km north of here, the M2 staurolite isograd trends north-south and is truncated by M3 isograds, suggesting that this locality may have experienced M2 previously. This interpretation is also suggested by small euhedral staurolites within micaceous pseudomorphs in the M3 staurolite zone northwest of here. Continue east and then northeast on US-2 and ME-4.
- 53.1 STOP 6. Park on right. Extensive exposures on right are staurolite-zone rocks of the Sangerville Formation. Staurolite-biotite-garnet schist (4A) is common. As is the case elsewhere in this quadrangle, andalusite has not been seen. A typical 4A specimen has staurolite with Mg# 15.9 and biotite with Mg# 46.9. Some specimens contain pods of medium-grained muscovite similar to spangles but smaller. These may be related to previous M2 effects which were garnet-grade in this area. The exposure shows significant alteration, and greener

portions are extensively altered to chlorite, perhaps by the Hercynian (M5) metamorphism which almost reached the southwest corner of Farmington quadrangle (Fig. 2).

- 53.5 Extensive roadcuts on right of staurolite-biotite-garnet schist (4A).
- 53.7 Roadcuts on left and right are mica schists, biotite quartzites, and calcareous rocks.
- 56.1-56.4 Sparce roadcuts of staurolite-biotite-garnet schist (4A).
- 57.6 Roadcuts on the southwest side of Farmington are garnet-biotite-chlorite schists (3) and biotite schists. These rocks were in the garnet zone of both M2 and M3. They show two distinct stages of growth with an abrupt transition from Mn-Ca-rich cores to Fe-rich rims.
- 58.1 Cross Sandy River on US-2 and ME-4 in Farmington.
- 58.3 Take right fork southeast on US-2 and ME-27.
- 62.1 Roadcuts on left of vesuvianite-bearing calc-schist in calcareous horizons of the Sangerville Formation, and granitic dike.
- 62.7 Turn right on ME-41-156 to Farmington Falls.
- 63.2 STOP 7. Park on either side of road before bridge. Outcrops under bridge and west of bridge are staurolite-zone rocks of the Sangerville Formation. Rocks are staurolite-biotite-garnet schists (4A) with idioblastic staurolite, generally with two stages of growth and a chemical break between the main crystal and the outer rim. From two specimens, staurolite has Mg# 13.6 and 14.1, and biotite has Mg# 48.0 and 45.0. About 1 km south of here are plutonic rocks (Fig. 6) of the Livermore Falls group.

Although andalusite is almost totally absent from M3 rocks of the Farmington quadrangle, 14 km east-northeast of Farmington Falls at East Mercer, andalusite occurs and becomes common as does cordierite around the north side of the Skowhegan (or Rome, or Norridgewock) batholith. Staurolite is absent here. In addition, there is an eastward narrowing of metamorphic zones. These observations and absence of evidence of polymetamorphism suggest a decrease in P of M3 at its northeastern extreme, perhaps due to a decrease in surface volcanic activity during Acadian magmatism. Turn around and turn right toward New Sharon.

- 63.3 Rejoin US-2 and ME-27, turn right (east).



- 67.3 Turn left (north) on ME-134 right before bridge over Sandy River. Most of the route between here and Stop 8 is over chlorite- and biotite-zone rocks.
- 68.9 Enter Norridgewock quadrangle.
- 75.3 Outcrops in stream of quartzite and chlorite phyllite.
- 75.8 Turn right (east) on ME-43.
- 76.9 Roadcuts on left of biotite quartzite and biotite-chlorite phyllite.
- 79.9 Enter Anson quadrangle.
- 80.2 Roadcuts on right of Old Point pluton, a small northern satellite of the Norridgewock batholith.
- 82.7 ME-148 joins ME-43 from the left, proceed right on ME-43-148.
- 84.1 Without crossing bridge over Kennebec River, proceed straight (north) on US-201A and ME-8.
- 88.7 Extensive outcrops left of bridge over Carrabassett River are quartzite, biotite quartzite, and slightly calcareous rocks of the Fall Brook Formation.
- 95.6 Cross Kennebec River. The M2 staurolite isograd occurs in this area, trending northeast.
- 96.8 Turn left (north) on US-201 in Solon. Outcrops on right in stream are Madrid quartzites and biotite quartzites.
- 97.4 Turn left on Falls Rd.
- 97.7 Proceed straight on poorer road.
- 97.8 STOP 8. Park along dirt road to left. NOTE SPILLWAY SIGN. Arnolds Landing, staurolite-grade pelites of the Carrabassett Formation. Large porphyroblasts are retrograded andalusite, and smaller ones are partly retrograded staurolite and garnet. Assemblages include staurolite-biotite-garnet schist (4A), staurolite-andalusite-biotite-garnet schist (4B), and andalusite-biotite schist (4C). Retrograde chlorite is common and biotite is extensively altered. The bulk chemical control on mineral assemblages is obvious in the bed to bed variation. The 4A and 4B rocks are more pelitic in character and generally have smaller porphyroblasts. The 4C assemblage is seen in more quartz-rich horizons with coarse andalusites (up to 4 cm). Redox is not a factor in this variation as all oxides are hematite-free ilmenite, and graphite is ubiquitous. A

possible sedimentological explanation for these observations is increased kaolinite/illite ratio in the siltier layers. Chlorite-grade phyllites are exposed 2 km east-southeast on the east side of Solon. West of here about 6 km is a rapid increase in grade (M2s) near the Lexington batholith (Fig. 7).

The heat source for M2 is believed to have been a deeper, more extensive phase of the Lexington batholith which heated the overlying rocks. Contact metamorphism around the central and south lobe (M2s, Figs. 1, 7) slightly postdates regional M2. Return to US-201.

- 98.1 Turn left on US-201.
- 100.0 Enter Bingham quadrangle. (Note 45° N. Lat. sign. Elevation is 420 feet, not 352).
- 100.7 Road cuts on right of Madrid Psammitic rocks.
- 101.6 Road cuts of staurolite schist, Carrabassett Formation(?).
- 105.3 Turn left (west) across Kennebec River on ME-16 in Bingham.
- 105.5 Turn right (north) after bridge. Road cuts are Smalls Falls sulfidic-pelitic schists and quartzite. Andalusite is rare.
- 107.1-107.6 Muscovite-rich pelites of Perry Mountain Formation.
- 107.6-107.9 Smalls Falls Formation.
- 109.8 Take left fork toward Rowe Pond.
- 111.7 STOP 9. Park on right. Carrabassett pelites near the base of the staurolite zone. Here metamorphic grade increases south. Much of the roadcut contains staurolite without andalusite, but a 10 m zone near the middle also contains andalusite as slender 10 cm porphyroblasts. Assemblages are staurolite-biotite-garnet-chlorite (4A) and staurolite-andalusite-biotite-garnet-chlorite (4B). Some chlorite appears to be primary. Analyses of a 4A specimen give staurolite Mg# 11.9 and biotite Mg# 33.7. North-northwest 300 m staurolite and chloritoid coexist in rocks transitional to the M2 garnet zone. The Stop 9 locality may represent Reaction 2, with all reacting minerals present in some specimens. North-northwest 3 km are cordierite-bearing, staurolite-absent contact metamorphic rocks of M2n around the north lobe of the Lexington batholith, which are earlier (?) and formed at lower P (2.35 kbar). Proceed straight ahead.
- 112.3 Turn right down hill.



113.2 Turn right. Return to Farmington without crossing the Kennebec River via this road to ME-16-west, south to US-201A, south to North Anson. Here take ME-234 west from south of bridge in North Anson to ME-17, and ME-17 south through the New Vinyard Mountains to Farmington.

## Mineralogic and Textural Evidence for Polymetamorphism Along a Traverse from Oquossoc to Phillips to Weld, Maine

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### INTRODUCTION

This trip is aimed at examining some of the textural and mineralogic features that have resulted from polymetamorphism of the Cambrian through Devonian strata that underlie much of western Maine. Because these features are strongly influenced by the nature of the metamorphism which have affected the rocks, it will be necessary below to provide a brief review of the metamorphic styles that are present. Evidence will be examined for one Ordovician and several Devonian metamorphism. The latter occurred as part of the Acadian Orogeny.

Outcrops will be examined in portions of the Oquossoc, Rangeley, Phillips and Dixfield 15' quadrangles. In order to best accommodate with the field trip time constraints, we will begin the trip at the stop which is most distant from our Farmington base and work our way back to Farmington during the day.

By way of acknowledgements, it should be noted that most the author's petrologic work in Maine has been supported by various NSF grants since 1965. Support for the author's mapping of the Oquossoc quadrangle was provided by the Maine Geological Survey during the summers of 1962-64. In addition, the author would like to acknowledge his gratitude to Dr. R.H. Moench for providing detailed geologic base maps of the Rangeley and Phillips quadrangles such that the author then merely had to collect specimens in these areas. Moreover, Dr. Moench provide other valuable assistance for my collecting goals, including provision of a considerable number of samples from more remote areas.

### REGIONAL GEOLOGY

Geologic mapping of the four 15' quadrangles to be visited has been provided by Pankiowskyj (1964) (Dixfield), Guidotti (1977) (Oquossoc), Moench (1971) (Rangeley and Phillips). In addition to numerous topical geologic reports by Moench on western Maine, a broad synthesis which includes the four quadrangles of concern has been provided in Moench and Zartman (1976). One should also consult the new Geologic Map of Maine (Osberg et al. 1985) for a broader overview placing the geology



of western Maine into the context of the geology of the whole state. For our purposes the most important aspects of the geology in western Maine are:

(1) The stratified rocks range in age from Cambro-Ordovician to Devonian. Most of the pre-Silurian units occur in the northwestern part of the Rangeley quadrangle and in the Oquossoc quadrangle. By far the bulk of the stratified units consist of interbedded metapelites and impure metasandstone to greywacke. However, some units consisting of metabasalt, impure quartzites, and calcareous sandstones are also present. Nonetheless, pelitic bulk compositions are so abundant and widespread that the metamorphisms that have taken place can be studied in terms of this bulk composition.

(2) All of the strata have been intensely deformed into mainly, northeast-trending folds and faults. Some northwest-trending structures are also present but they are clearly subordinate. Most of the folding and faulting occurred before any of the high-grade metamorphic events. However, the deformation was probably accompanied by low-grade metamorphism throughout much of the area. Moreover, in the case of the Cambro-Ordovician units, deformation and low-grade metamorphism may have also occurred during pre-Silurian events (Harwood, 1973). Post-metamorphic structures include a few faults and in a few cases one finds minor microscopic evidence of post-metamorphic kinking of micaceous minerals.

(3) Various types of plutonic bodies are present with a range in sizes and shapes. Most of the smaller bodies occur as dikes and sills or have irregular shapes. Compositionally these are mainly granitic pegmatites and two-mica adamellites. As expected, they are most common in the areas of higher metamorphic grade.

The larger plutons are identified on Figure 2 and include:

(a) Mooselookmeguntic (371 Ma) and Phillips Plutons. According to Osberg et al. (1985) these bodies include lithologies ranging from two-mica granite to tomalite. Of particular importance is the fact that they occur as large, low-dipping sheets. For our purposes it should also be noted that they are intimately interrelated with most of the metamorphism we will be inspecting.

(b) Reddington Granite. This body consists of coarse-grained granite with distinctive feldspar phenocrysts. It does not include abundant binary granite phases. Moreover, the small amount of muscovite present may be due to later metamorphism. Inasmuch as the hornfels around the Reddington granite appears to have been affected by the  $M_3$  metamorphism which, it will be argued below, was caused by the intrusion of the (371 Ma) Mooselookmeguntic batholith, it would appear that the Reddington granite is a somewhat older Devonian pluton.

(c) Unbagog granodiorite. This body outcrops in the southwestern part of the Oquossoc quadrangle. It is not of importance for the goals of this field trip.

(d) Adamstown granite. This granite was first mapped as part of the study by Guidotti (1977). It is a member of the Highlandcroft magma series and has a crystallization age of  $452 \pm 4$  Ma (Lyons et al. 1986). For the purpose of this field trip it is of interest because it produced an Ordovician contact metamorphic aureole. Later, it was deformed during the Acadian Orogeny and also variably metamorphosed. The effects of the deformation and the variable degrees of annealing by the subsequent, static metamorphic events are readily observed even in outcrop.

### OVERVIEW OF METAMORPHISM

Polymetamorphism in western maine was first discussed by Pankwiskyj (1964). Later, Guidotti (1970a) provided more detailed documentation for two, post-tectonic, high-grade Devonian metamorphism in portions of the Rangeley and Oquossoc 15' quadrangles. These two events were superimposed on an earlier low-grade Devonian metamorphism. Subsequently, multiple Devonian metamorphic events were recognized elsewhere in western and central Maine, (e.g. Henry (1974), Cheney (1975), Novak and Holdaway (1981) and adjacent New Hampshire (Wall, 1988).

Much of this work was drawn into a synthesis by Holdaway et al. (1982). They described, in a general way, a total of four metamorphic events.

$M_1$  - a low-grade event which was probably syntectonic.

$M_2$  - a regionally extensive, post-tectonic event which in wide areas resulted in the establishment of the AFM join connecting andalusite and biotite. This event affected a wide area in western and central Maine.

$M_3$  - a regionally extensive, post-tectonic event which over a broad area resulted in the establishment of the AFM join connecting sillimanite and biotite. As with  $M_2$ , this event affected much of western and central Maine.

$M_4$  - a high-grade, low pressure metamorphism, closely concentrated around plutons such as the Reddington granite.

The first three events were recognized as being essentially the same as those recognized by Guidotti (1970a). The fourth event,  $M_4$ , has subsequently been found to be earlier than  $M_3$  and possibly  $M_2$  -- at least in the Rangeley-Phillips areas where it is well developed as a contact-hornfels around the Reddington granite.



Holdaway et al. (1982) emphasized that  $M_2$  and  $M_3$  may not have been completely synchronous throughout the whole area they considered. They also suggested that a close genetic relationship existed between the  $M_3$  isograds and large sheet-like plutons such as the Mooselookmeguntic batholith. However, they stopped short of attributing the  $M_3$  isograds directly to the heat imparted by the intruding plutons.<sup>3</sup> The distribution of  $M_3$  metamorphic grades as well as the gross distribution of the high-grade portion of  $M_2$  have been shown by Guidotti (1985) on the new Geologic Map of Maine.

Subsequent work (e.g. Holdaway et al. (1986) (1988), Lux and Guidotti (1985) has modified the suggested distribution of the  $M_2$  and  $M_3$  isograds and recognized the existence of an even later Carboniferous event. In addition, combining the petrologic studies with  $^{40}\text{Ar}/^{39}\text{Ar}$  work and thermal modelling has enabled Lux et al. (1986), Guidotti et al. (1986), DeYoreo et al. (1989A,B) to develop an integrated model for the interrelationship between the plutons and metamorphism in western Maine. Briefly the model suggested includes:

(1) The post-tectonic high grade metamorphism are directly caused by heat convected in by the emplacement of sheet-like plutons. Basically the metamorphism involves shallow to deep level contact metamorphism (see Fig. 1).

(2) The extensive areal development of these contact events can be attributed to the shallow dips of the sheet-like plutons that brought in the heat.

(3) Inasmuch as heating is directly tied to pluton emplacement, the lack of rigorous, regional synchronicity of  $M_2$  or  $M_3$  is made understandable. Moreover, the post-tectonic aspect of virtually all of the high-grade metamorphisms in western Maine is also explained by this means.

(4) An important aspect of the model developed is that it involves short-lived, thermal pluses, each one being essentially isobaric. As seen on Fig. 1, the various metamorphic paths suggested occur over a range of pressures, but the vast majority involve pressures less than that of the Al-silicate triple point (e.g. see also Tewhey (1975), Holdaway et al. (1986, 1988), Cheney (1975), Henry (1974)).

(5) The model argues for cooling to ambient temperatures before each succeeding thermal pulse. As a result, where the metamorphic pulses overlap spatially it has resulted in polymetamorphism. In some cases an area affected by an earlier metamorphism will be prograded in one portion and downgraded in another. As will be seen on the field trip, this aspect, plus the static nature of the heating events, has resulted in extreme development of pseudomorphs -- either prograde or downgrade as the case may be.

Most of the above discussion has been directed at the events,  $M_2$  and  $M_3$ , which in much of western Maine attained high-grades. It was implied above that these events were statically superimposed on an earlier, greenschist grade  $M_1$  event. The evidence for such a superposition can be seen best in thin section although sometimes evident megascopically in outcrop.

In thin section one commonly observes that the earliest foliation has been cut by a later slip cleavage. Subsequently, both of these cleavages have been helicitically overprinted and truncated by both staurolite and andalusite and in some cases by biotite and chlorite plates. The point to note is that these observations indicate development of a micaceous foliation before any of the later events which attained high grades. This implies at least some early recrystallization of these rocks -- probably syntectonically. It is this early recrystallization which is designated as  $M_1$ .

As seen on Fig. 2, the area around the village of Rangeley and to the north and west shows no evidence of ever having been affected by metamorphic grades exceeding garnet zone or greenschist facies. It is quite possible that much of that area has been affected by all three ( $M_1$ ,  $M_2$ , and  $M_3$ ) Devonian metamorphism. Unfortunately, greenschist grade superimposed upon greenschist grade leaves little evidence of more than one event -- especially if the later events involved static recrystallization. Indeed as suggested above, some of the Cambro-Ordovician units we'll see in the early part of the field trip may have experienced an even earlier, pre-Silurian greenschist facies metamorphism.

Only in one case is there clear evidence for a pre-Silurian metamorphism. It will be seen at Stop 2 as a "fossil" hornfels which developed around the Ordovician, Adamstown granite ( $452 \pm 4$  Ma, Lyons et al. 1986). In turn, we will also be able to see clear evidence that this granite was deformed and metamorphosed by subsequent Devonian events.

For the purposes of this trip, the above general review should suffice. Any additional required elaborations will be given in the context of the specific field trip stops. Fig. 2 shows the areal distribution of the  $M_3$  isograds in the portion of western Maine relevant to our needs. Also shown is the approximate trace of the  $M_2$  isograd marking the establishment of the AFM join connecting biotite and andalusite (or staurolite).

Guidotti (1989) has presented a more complete review of the nature of the metamorphism that have affected western Maine. However, for details one should consult the papers by Holdaway et al. (1986) (1988), Lux and Guidotti (1985), Lux et al. (1986), DeYoreo et al. (1989A)(1989B), and Guidotti et al. (1986).



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## ITINERARY

This trip will meet in Farmington at MacDonalds. From there we will drive northward to the first stop which is near the north-central border of the Oquossoc 15' quadrangle. We will assemble at MacDonalds and drive northward on Rte 4, passing through Rangeley. At the east edge of the village of Oquossoc, turn right (north) on Rte 16 and loop around the north end of Cupsuptic Lake. After the loop, the first stop is at the top of the first large hill.

It would be well to bring lunch with you although in mid to late morning we'll be passing through Rangeley and you could pick up lunch material at Stubby's market. 15' topographic maps that will be useful are: Oquossoc, Rangeley, Phillips, and Dixfield. However, the road log is based upon the 2° Lewiston sheet.

### Milage

0.0 STOP 1: Over the last mile we have been driving through outcrops of the Adamstown granite, an Ordovician, Highlandcroft Magma Series pluton. All of these outcrops show significant amounts of deformation and low-grade metamorphism of the original granite. This deformation and metamorphism are due to effects of the Acadian Orogeny. As seen on Fig. 2 one can trace the Devonian metamorphic isograds through the Adamstown Granite -- at least in a rough fashion. Here, at Stop 1 we are in the low grades of the Devonian metamorphism.

At this stop one can see a saussuritized plagioclase groundmass enclosing quartz lenses and K-spar augen. Note on the weathered surfaces how the quartz lenses do not form an interlocking texture. In thin section one finds the quartz lenses to be surrounded by fine-grained granular quartz grains.

At these low grades the biotite has been altered to chlorite. Some newly formed epidote can be seen even in hand specimen whereas metamorphic muscovite can be seen only in thin section. Typically the K-spar megacrysts are pink in this low grade portion of the pluton.

You may find it useful to collect a small piece of this low-grade meta-granite so you can compare it to specimens seen at Stop 4 where the metamorphic grade is in the lower sillimanite zone.

Return to cars and continue on Rte 16.

1.0 Turn right on old paved road - this is the old location of Rte 16. It is drivable but requires considerable care to avoid holes etc.

1.75 Adamstown granite on the right.



- 1.9 STOP 2: Here we see several, scattered outcrops of a dense, massive, hornfelsic rock that has a hackly fracture. We are essentially on the contact with the Adamstown granite and it caused a narrow hornfels to develop in the Cambo-Ordovician Aziscohos fm. Hence, we are seeing the effects of an Ordovician metamorphism upon which later, low-grade Devonian metamorphism has been superimposed.

Look closely and you can find the thin laminae typical of the Aziscohos fm. On some of the glacially smooth surfaces you can find light colored patches which are pseudomorphs after highest grade minerals which formed during the Ordovician metamorphism. Hence, we are looking at a "fossil" hornfels.

Return to cars and continue in same direction.

- 2.2 More outcrops of the hornfels. Some may involve metavolcanics.
- 3.05 The washed out gully on the left has good exposures of the more typical greenschist grade Aziscohos fm. We are now out of the "fossil" hornfels and one finds green phyllites with quartz stringers.
- 3.87 Intersection with the new Rte 16 - turn left.
- 4.10 Very large outcrops of the phyllites.
- 4.45 Adamstown-Cupsuptic town line.
- 5.45 Massive greenstones and some phyllites.
- 5.60 Green phyllites with abundant, vein quartz stringers.
- 6.55 Rangeley-Cupsuptic town line and Oxford-Franklin county line.
- 9.80 Outcrops of massive metavolcanics (greenstones).
- 10.10 More outcrops of massive metavolcanics (greenstones).
- 10.20 Outcrops of foliated to massive metavolcanics (greenstones). Some involve very coarse-grained material as well as epidote-rich patches.
- 10.55 Intersection with Rte 4 - Turn right (west).
- 10.90 Downtown Oquossoc and the Oquossoc Hotel.
- 11.85 STOP 3: Park and walk back to the outcrop. For the cars at the back of the line, note that a public parking area

is available on the left, just across the road from the outcrops.

Here we will see the green phyllites of the Cambro-Ordovician Dead River fm. The Acadian metamorphism here are about at biotite zone.

Two things to note are the occurrence of abundant quartz stringers and magnetite octahedra. The origin of the quartz stringers can be debated, but the point of interest here is that when seen at higher grades (next Stop) one could mistake them for migmatites.

The magnetite is of interest because its presence is typical of many of the Cambro-Ordovician units in western Maine whereas few of the Siluro-Devonian units seem to have any magnetite present. Work by the author and M. Darby Dyar (University of Oregon) has shown that the biotite coexisting with magnetite in these units typically has about 20% of its Fe as  $\text{Fe}^{3+}$  whereas the biotites of the Siluro-Devonian units (typically containing graphite) have only 10% of their total Fe as  $\text{Fe}^{3+}$ . Moreover, only in the former is there significant VI  $\text{Fe}^{3+}$ . Probably reflecting these differences, one finds that biotites from the Cambro-Ordovician units has a dark, greenish-brown color. Biotite from the Siluro-Devonian units typically occurs in shades of orange to dark reddish brown, depending upon the metamorphic grade-controlled Ti content.

11.95 Turn left (south) on Bald Mtn. road.

14.00 STOP 4: Park carefully on both sides of the road and walk back to the outcrops. This outcrop was Stop 2 of Guidotti (1970a). Here, we see three rock types - all in lower sillimanite zone or upper staurolite zone of  $M_3$ . Possibly these rocks were also affected by  $M_2$  high-grade metamorphism although Fig. 2. indicates otherwise. Insufficient detailed work has been done to answer this latter question.

(1) The phyllites containing quartz stringers which we saw at the last stop are now dark-grey metapelites due to abundant biotite and only minor chlorite remaining. In some cases, thin section study shows development of some late, coarse clots of chlorite after biotite. Such chlorite is especially common in the NW corner of the Oquossoc quadrangle and in the Errol quadrangle to the west. Its origin is not yet understood.

At the present outcrop one can find good  $1/4''$ , euhedral garnets and tourmaline; occasionally staurolite is also seen in hand specimens. Sillimanite is always



microscopic. Note how the quartz stringers simulate a migmatite.

(2) Some metavolcanics are common as interbeds in the Cambro-Ordovician units. While driving, we saw them as greenstones in the lower metamorphic grades. Here they are dark amphibolites reflecting that higher grade has destroyed the epidote, actinolite, albite etc mineralogy and produced a rock composed largely of hornblende, calcic plagioclase and quartz.

(3) Here we can also see good outcrops of the Adamstown granite where it has been affected by the post-tectonic, high-grade  $M_3$  metamorphism. Although it has some development of a foliation here, in comparison with what we saw at Stop 1 it looks much more like a normal granite. This is probably a reflection of metamorphism at high grade. Moreover, the clastic texture so evident at Stop 1 is now absent.

Return to cars and continue southward.

- 14.23 Make a U-turn at the junction with the woods road coming in from the left and head back to the north.
- 16.50 Stop Sign: Turn right on Rte 4 heading to the east toward Oquossoc village.
- 17.50 Oquossoc village again - continue to the east on Rte 4, past the junction with Rte 16.
- 18.00 STOP 5: Metavolcanics on both sides of the road. This was Stop 1 of Guidotti (1970a). These outcrops belong to the Ordovician Kamankeag fm.

The rocks are typical greenstones (nearby pelites indicate biotite zone) containing actinolite-chlorite-epidote-albite and some calcite, quartz, sulfides, etc. The protolith is presumed to be basaltic flows but some pyroclastics are also present in other outcrops.

Notice the epidote-rich pods and clots in the otherwise massive to foliated volcanics. Possibly these represent volcanic bombs.

In some cases, remnant plagioclase and pyroxene grains have been found in these rocks.

- 18.7 Outcrops of the Kamankeag slates. About four miles northward along strike these rocks contain graptolites, Berry and Harvard (1967).

Continue eastward through Rangeley. For those who need to buy lunch materials, Rangeley will be your last chance.

26.00 At the Farm House Inn on the east side of Rangeley Village. We've been driving past many outcrops of the Quimby fm. All are at low grades - biotite zone at most. For our purposes, the key point is that there is no evidence of any sort that these rocks were ever affected by higher grades, nor any evidence as to how many metamorphic events have affected them.

26.43 Town line of Rangeley and Rangeley Plantation.

26.77 STOP 6: Outcrop across the road from the Terraces. Park on wide shoulders. This is the Greenvale Cove fm. It consists of fine-grained, thin bedded slates. Tiny biotite tablets are visible in hand specimen.

As with all of the outcrops since Stop 5, there is no sign of garnet or any kind of pseudomorphs in the metapelites. Hence, the rocks at this outcrop have probably never been at any higher grade than biotite zone. Some thin calc-silicate beds at the lower portion of the outcrop do contain garnet but none occurs in the metapelites.

At the lowermost outcrops one gets into grits of the Rangeley fm.

27.00 Grits of the Rangeley fm - also Town line.

27.12 Large crops of the cgl facies of the Rangeley fm.

27.82 Bear left (carefully) onto old Rte 4.

27.97 STOP 7: At Greenvale School on the old 15' quadrangle (now the Town Office of Sandy River Plantation). There is plenty of parking here so please take care not to block peoples driveways.

Walk up the old road in back of the Town Office and into the small quarry on the power line. The rocks here are metapelites of the Rangeley fm. Some calc-silicate nodules are also present.

The key thing to note are the excellent, euhedral pseudomorphs of staurolite. Garnet also appears to be wholly pseudomorphed. Can anyone find any fresh (remnant) garnet? You will note that the orientation of the original staurolite seems to be nearly perpendicular to bedding.



Head up along the path parallel to the brook to Cascade Falls. As you walk note the loose blocks along the trail as they show spectacular development of the euhedral pseudomorphs after staurolite (PLEASE DON'T HAMMER).

On the brook you can inspect the large faces containing abundant pseudomorphs. Here you can hammer at will! Can anyone find evidence of pseudomorphs after andalusite? To this point I've found only the staurolite pseudomorphs.

Hence, I interpret these rocks as having attained only staurolite grade during  $M_2$  but having been downgraded to biotite zone by  $M_3$ . Inspection of the spatial relationships between the pseudomorphs and groundmass foliations at many outcrops suggests that both  $M_2$  and  $M_3$  were largely static recrystallizations.

Return to cars and continue in same direction.

28.02 Encounter new Rte 4 again - bear left.

29.42 Road to Long Pond (Edelheid Road).

30.32 On the left - fairly rusty weathering Rangeley schist. Probably more recrystallized than at Stop 7 as indicated by the groundmass being a "white schist". This is a reflection of the groundmass muscovite being more coarse grained. Pseudomorphs can be found in the more pelitic beds and some garnet seems to be present although it is very sparse due to the sulfitic nature of this rock.

31.42 STOP 8. Large outcrops of Rangeley fm on the right. Here, there is a great display of pseudomorphs in a groundmass that is generally coarser grained than at Stop 7.

I've shown this on Fig. 2 as being in the  $M_3$  garnet zone but garnets are hard to find due in part to the moderately sulfitic nature of the outcrops.

Can anyone find "live garnet" or pseudomorphs after andalusite?

This is a good outcrop for collecting.

36.03 Town line between Sandy River Plantation and Letter E Township. We have been driving by many outcrops of Rangeley fm. In most of them there is development of euhedral pseudomorphs of staurolite. However, in some of the outcrops where the Appalachian trail crosses Rte 4 one can find samples with staurolite ranging from fresh to wholly pseudomorphed. This is a localized zone

forming on "island" in a broad area within which the staurolite has been totally pseudomorphed. In a later stop we will inspect some similar outcrops.

- 36.20 STOP 9. On the sharp, hairpin curve. Good parking at and beyond the curve but use extreme care watching for cars. Go to the lower grey outcrops. This is the Perry Mtn fm. and we are just below the type locality.

Notice the well preserved bedding features; cross beds, graded beds etc. The main foliation is essentially parallel to bedding but a slip cleavage is also present and shows up in outcrop as a pronounced crinkling.

The groundmass of the pelitic beds is a nice "white schist" due to recrystallized muscovite. Garnet is partially replaced by coarse chlorite but minor "live garnet" has been observed. Hence, these rocks are shown on Fig. 2 as  $M_3$  garnet zone.

The staurolite pseudomorphs are well developed and obvious - but can you find the pseudomorphs after andalusite?

Return to cars and continue down hill.

- 37.85 STOP 10. Turn right into the Smalls Falls picnic area and drive to the end of the parking area.

This is the type locality of the Smalls Falls fm. From a petrologic view point it is an interesting unit because it contains 5-10 modal % of pyrrhotite and also is graphite-rich. As a consequence it shows the effects of sulfide-silicate mineral reactions to an extreme degree. The most straight forward aspect of these reactions is that the silicate bulk composition is moved to a very Mg-rich portion of composition space. Hence, at the appropriate grades one finds assemblages like andalusite or sillimanite + Mg-cordierite + phlogopite. Fe-rich minerals like garnet or staurolite never occur in this unit. Moreover, the Ti phase is rutile instead of ilmenite as in the other Siluro-Devonian metapelites.

Walk to the outcrop at the edge of the plunge pool. On the smoothed surfaces you will find large chiastolite crystals -- except that, as shown in Guidotti and Cheney (1975) they are now pseudomorphs composed mainly of margarite. As your eyes become more focused on the details of the textures present, you will see that other "1" pseudomorphic knots are present. They consist of aggregates of chlorite and phlogopite after cordierite. The original chiastolite and cordierite formed during  $M_2$  but during  $M_3$  they have been pseudomorphed.



Please don't hammer on the outcrops in the park area. The best opportunity for collecting margarite is at the road outcrops just uphill from where the cars are parked.

Return to cars and drive back to Rte. 4.

38.10 Rte 4. Turn right (south).

41.00 STOP 11. Downtown Madrid - Walk to outcrop on the river by the bridge across the road coming in from the east.

This is the type locality of the Madrid fm which consists of an interbedding of dense biotite granulite and calc-silicate.

For our purposes note the nice examples of bulk composition control of mineralogy. In particular, note that a few pelitic beds are present and they display good pseudomorphs after staurolite. Some of the pseudomorphs appear elongate parallel to a slip cleavage so that they look deformed. However, others are clearly undeformed. The elongation may reflect growth of the original staurolite along the direction of the slip cleavage.

In a few cases, the shapes of the pseudomorphs make one speculate if they were originally andalusite.

Some of the calc-silicate beds show evidence of metasomatic zoning suggesting some degree of diffusion across strong compositional gradients. Note the slightly purplish cast to the biotites in the beds associated with the calc-silicates. Many people have noted that cast for biotites of calcareous rocks, but I am not aware of any explanations that have been advanced for this color effect.

Finally, some pegmatite veins and adamellite veins are present in places along these extensive outcrops.

Return to cars and continue southward.

41.90 STOP 12. Park just before the bridge. NO SMOKING at this outcrop at the request of the owner.

Walk down along river for about 50 yards and one encounters extensive outcrops of the Devonian, Carrabasset fm.

The point of this outcrop is that it is in one of those "islands" within the  $M_3$  garnet zone in which  $M_2$  staurolite is variably pseudomorphed. Indeed, with a little searching you can also find variably pseudomorphed andalusite at these outcrops.

Focus on the nature of the staurolite pseudomorphs in some detail. Here, they are formed as the result of downgrade effects. At Stop 14 we'll see pseudomorphs of staurolite resulting from prograde effects.

Return to cars and head to the south.

53.00 STOP 13. Park carefully on the right taking care not to get stuck in soft sand. Watch out for cars.

On the left is a large outcrop of Carrabasset fm with perfectly fresh staurolite. It contains St + Bio + gn + chrt and so is a typical staurolite-grade assemblage for this part of western Maine. However, it appears to be another one of these "islands" of  $M_2$  staurolite that has withstood the  $M_3$  downgrading. This could reflect inability of  $H_2O$  entering some areas so that the downgrading was not effected.

One other possibility is that the areas in which  $M_2$  staurolite is preserved represents hot spots. This suggestion is based upon the geophysical study of Carnese (1983) which indicates that the sheet-like Mooselookmeguntic pluton extends beneath this whole area, reaching as far to the northeast as the Reddington Pluton. His cross sections suggest that it is not far below the surface such that local apophyses to higher levels may have produced  $M_3$  temperatures still within the staurolite stability field.

Obviously, resolution of such a suggestion will require a lot of detailed petrologic study.

Back to the cars and continue to the south.

54.13 Outcrops of Smalls Falls fm.

56.20 Turn to the right (south) on Rte 142.

58.50 STOP 14. Just north of the 1230 point shown on Rte 142 in the 15' Phillips sheet.

Coarse-grained, lower sillimanite zone metapelites. Fibrolitic sillimanite is visible in biotite.

Note the large aggregates of coarse muscovite and some biotite. Some of these aggregates contain remnants of fresh staurolite as well as abundant fibrolitic sillimanite in the biotitic outer rims of the pseudomorphs.

The concentration of coarse muscovite with some garnets in the pseudomorphs is very similar to the



prograde pseudomorphs reported in the Oquossoc area, Guidotti (1968). This coarse grained aspect plus the presence of the garnets seems to be typical of staurolite that has been pseudomorphed in a prograde fashion. You can contrast it with the numerous examples of downgrade pseudomorphs we've seen previously on this field trip.

Return to cars and continue southward.

- 66.00 Weld Corner - continue southward on Rte 142.
- 68.35 Stop sign in Weld village - continue straight across the intersection onto Rte 156.
- 71.10 Outcrops of the Phillips Pluton.
- 71.55 Calc-silicates, marbles etc.
- 72.15 STOP 15. Dangerous soft shoulders. Take care not to get stuck in the sand.

This outcrop is in the upper sillimanite zone as the metamorphic grade has exceeded the stability limit of staurolite.

Good development of fibrolite and small spangles of muscovite can be observed in the western most outcrop.

The main point of this outcrop is to discuss the occurrence of a late chlorite-forming event that can be detected in these rocks. Henry (1974) studied in some detail the petrology of the rocks in the NE 1/4 of the Dixfield 15' quadrangle. In most respects they are quite similar to the polymetamorphism described by Guidotti (1970A,B) (1974) in the Rangeley and Oquossoc areas.

However, it appears that a late, low T event has affected these rocks and produced a new generation of chlorite. In some cases it is clearly an alteration after biotite but in other cases the chlorite occurs as clean laths and plates. Henry (1974) discussed the pros and cons of whether it was in equilibrium with the rest of the assemblage in a given rock. It can be suggested that much of it may not be in equilibrium, mainly on the basis of three observations: (a) Some of this chlorite occurs in the upper sillimanite zone - something that does not occur in other parts of western Maine. (b) It does not show the compositional variation expected as grade changes (Teichmann, 1988). (c) In the staurolite and sillimanite zones the chlorite has a relatively Fe-rich composition (as evident even from its optical properties - purple brown colors in crossed nicols).

In addition to the questions raised by the nature of the chlorite, careful inspection of Henry's data for biotite and muscovite also suggests some anomalous features. The net assessment one arrives at is that subsequent to the  $M_2$  and  $M_3$  events, the rocks in much of Henry's study area have been affected by still another event, but one which attained only low temperatures. The cause and extent of this event can only be speculated about at this point. For example, it may represent an effect reflecting the northernmost extent of the Carboniferous event believed to have affected the southern portion of the Dixfield 15' quadrangle, Lux and Guidotti (1985).

Return to cars and continue southward.

- 81.53 Stop sign in town of Wilton at intersection. Turn left (east) and continue straight on this road (not Rte 156). It will take you to Rtes 2 and 4. Then go east to Farmington.



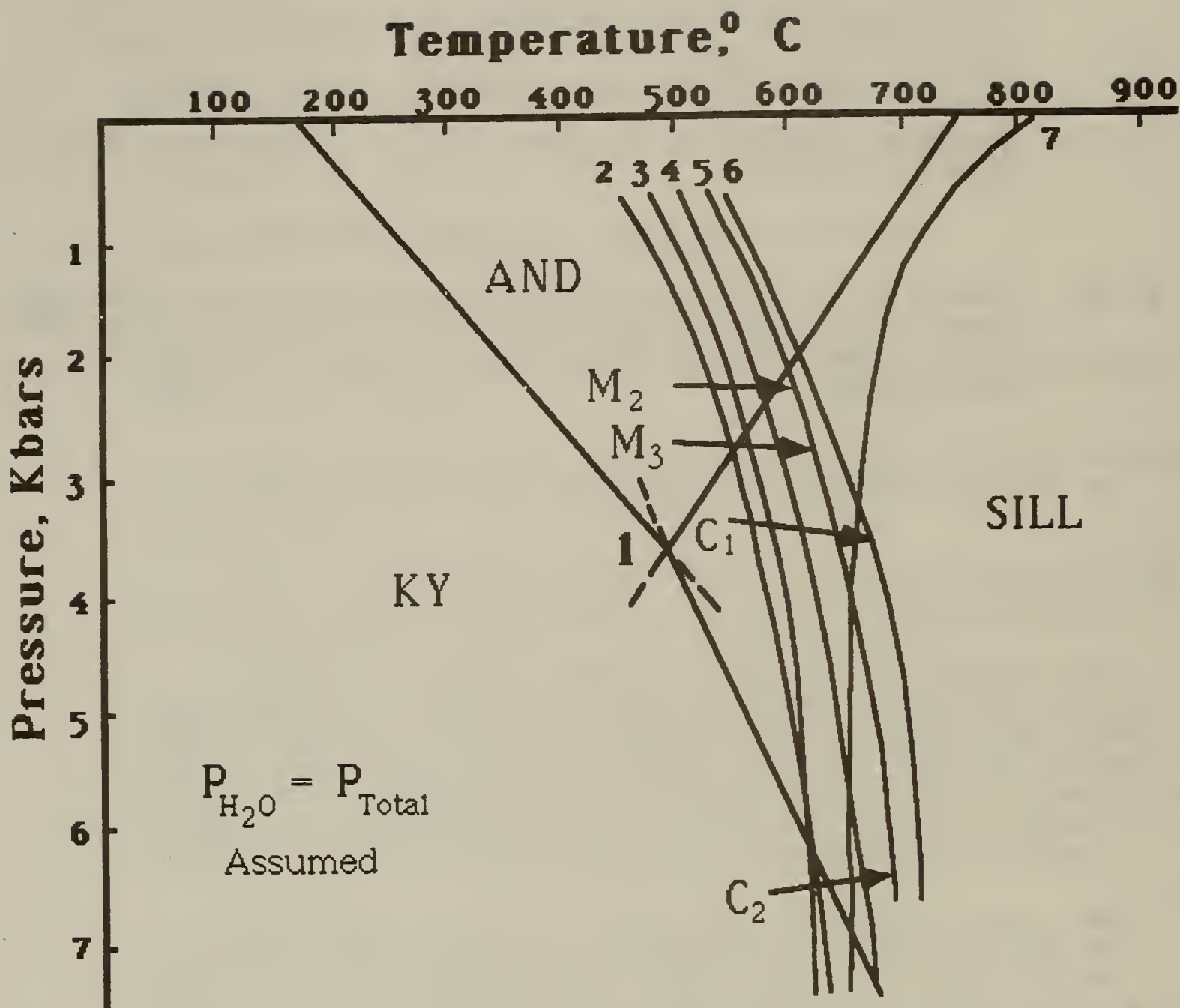
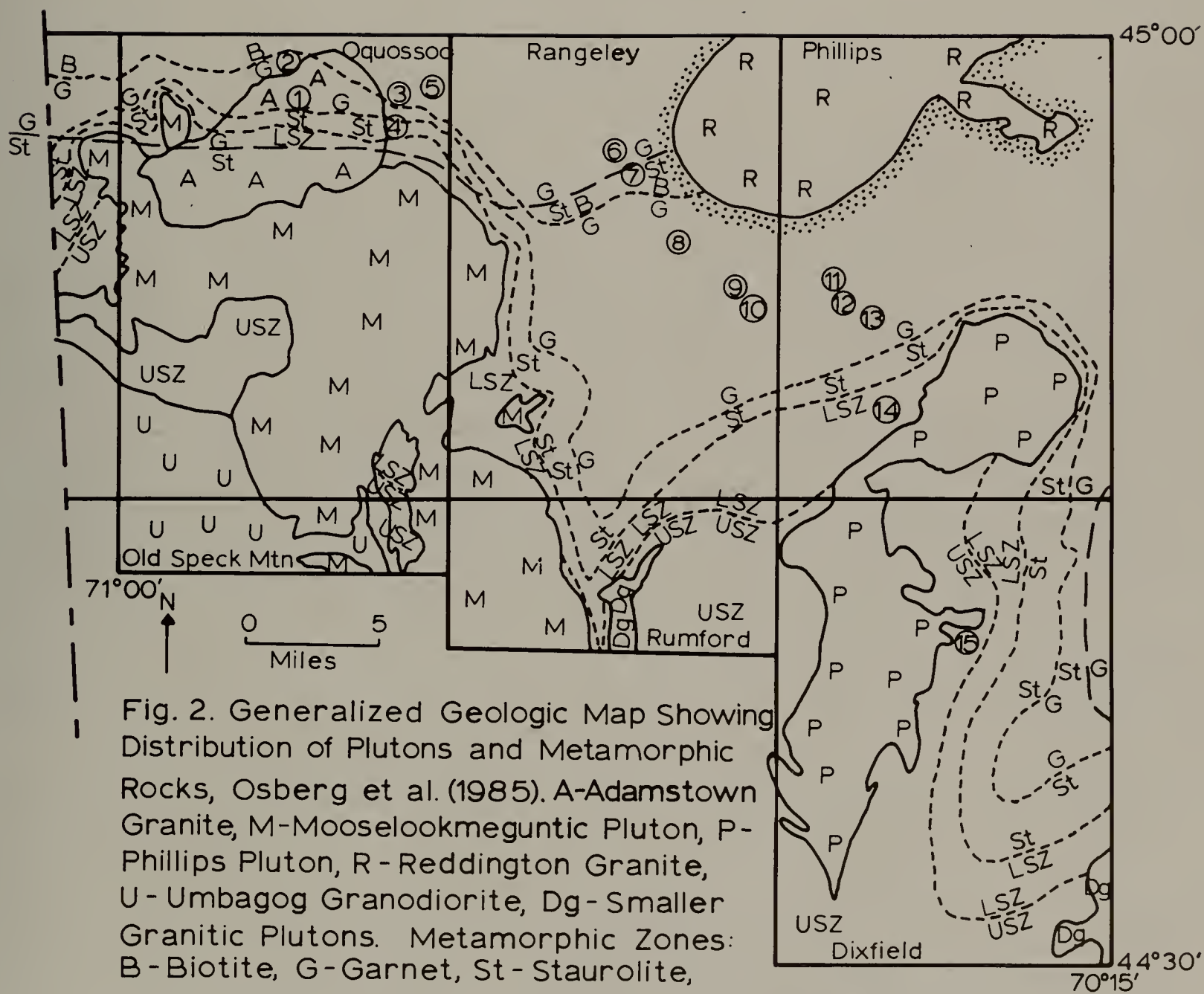


FIG. 1. PT GRID OF EQUILIBRIA RELEVANT TO THIS REPORT. (1) AL-SILICATE CURVES, HOLDAWAY (1971). (2) PARAGONITE BREAKDOWN, CHATTERJEE (1972). (3) ST + CHL = AL-SIL + BIO, GUIDOTTI (1974). (4) ST = AL-SIL + BIO, HOSCHEK (1969). (5) AB + MU = AL-SIL+KSP, CHATTERJEE AND FROESE (1975). (6) MUSCOVITE BREAKDOWN, CHATTERJEE AND JOHANNES (1974). (7) GRANITE MINIMUM, TUTTLE AND BOWEN (1958). M<sub>2</sub>, M<sub>3</sub> - PT PATHS OF FIRST AND SECOND HIGH-GRADE ACADIAN METAMORPHISMS RESPECTIVELY, C<sub>1</sub>, C<sub>2</sub> - PT PATHS OF CARBONIFEROUS METAMORPHISMS ON THE UPPER AND LOWER SIDES OF THE SEBAGO PLUTON RESPECTIVELY.







GLACIAL GEOLOGY OF THE ANDROSCOGGIN RIVER VALLEY  
IN OXFORD COUNTY, WESTERN MAINE

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## INTRODUCTION

This field trip will examine glacial deposits in a section of the Androscoggin River valley from Peru west to Newry, and thence up the Bear River valley to Grafton Notch in the Mahoosuc Range. The sites visited during the trip have been selected to illustrate the style of deglaciation in this part of Maine, and the variety of meltwater deposits resulting from retreat of the late Wisconsin ice sheet. Ongoing geologic mapping in the region has raised numerous questions concerning these deposits, which hopefully will stimulate discussion during the NEIGC conference.

The field trip area is located in the hilly to mountainous terrain of western Maine. Elevations are generally higher toward the northwest, culminating in the Mahoosuc Range where some mountains exceed 1,000 m. The Androscoggin River follows a peculiar zig-zag course as it flows eastward through the mountains. It is very narrow in this reach, and abruptly drops about 60 m at the falls in Rumford. Crosby (1922) discussed this "disarranged" drainage pattern. On the basis of geomorphic evidence, he proposed that the Androscoggin formerly flowed southward from Bethel along the present course of the Crooked River valley. Deposition of glacial sediments just south of Bethel supposedly blocked this course and diverted the river.

The products of deglaciation in the study area differ from those found in certain other parts of New England. The region is situated above the limit of late-glacial marine submergence, and lacks the abundant moraine ridges that occur in Maine's coastal lowland. Sand and gravel deposits formed by glacial meltwater are unevenly distributed along this section of the Androscoggin Valley, due in part to the valley's irregular east-west course. Esker systems enter the valley from tributary valleys to the northwest, follow the Androscoggin for a short distance, and then depart to resume their southward course through the hills in response to the former pressure gradient in subglacial tunnels.

Other meltwater deposits along the main Androscoggin Valley include glaciolacustrine deltas and fans, and glaciofluvial outwash. Base-level controls for the deltas and outwash are difficult to locate because of the erratic distribution of these deposits and scarcity of diagnostic exposures. The postglacial river has terraced the original deposits in some places and completely removed them in others. Deltas and subsurface records of lake-bottom sediments indicate that a glacial lake existed in Bethel (just upvalley from the area covered by this trip). Data are being collected to learn the



extent of this lake and others that probably formed in the Rumford area. Some of these lakes were dammed by ice and/or drift barriers in constricted parts of the Androscoggin Valley.

Positions of the retreating glacier margin are likewise poorly defined. There are few, if any, end moraines in the field trip area; and distinct ice-contact deltas or other deposits showing heads-of-outwash (like those which occur abundantly in southern New England) are relatively uncommon in this part of the Androscoggin River basin. However, the mapped deposits and meltwater channels indicate northward to westward ice recession over most of the area. Possible evidence for northeastward retreat of an ice mass in the vicinity of Rumford will be discussed during the trip. The timing of deglaciation in Maine is constrained by only a few key radiocarbon dates (e.g. Davis and Jacobson, 1985), from which it is inferred that the mountains in western Maine were uncovered between 14,000 and 13,000 yr B.P. (Thompson and Fowler, in press).

The topography of the study area is believed to have strongly influenced the pattern and mode of deglaciation. Southeast of this area, the stratigraphy and structure of end moraines indicate the presence of an active ice margin in the coastal lowland. To the west, the Androscoggin Moraine and other features provide evidence of active ice in the upper Androscoggin basin following emergence of the Presidential, Carter, and Mahoosuc Ranges from the ice sheet (Thompson and Fowler, in press). However, it is likely that large masses of ice were cut off from the thinning ice sheet in the lee of the Mahoosucs, in the central Androscoggin Valley and its tributaries. This local stagnation may account for the small volume of meltwater deposits generated in the latter region. The size and continuity of residual ice masses, and the extent to which they maintained some activity, are being investigated in conjunction with mapping studies in this area.

## DESCRIPTION OF STOPS

### STOP 1: West Peru Sections

This is a two-part stop. Stop 1-A is the large pit close to Route 108; Stop 1-B includes the small pit and washout gully along the logging trail leading to the south (other side of small intermittent stream) (Fig. 1).

STOP 1-A presently exposes about 3 m of section. Three stratigraphic units occur in this pit. From oldest to youngest, they are as follows:

(1) The lowest exposed unit is a sandy diamicton that is interpreted as glacial till, though its depositional environment is uncertain. The till is light olive-gray (5Y-6/2), non-oxidized, and variably stony. It contains assorted bedrock lithologies as angular to sub-rounded clasts to 1 m in diameter. Only a small percentage of these clasts are striated and faceted. Unit 1 closely resembles the typical surface till deposited by the late Wisconsin ice sheet in southwestern Maine (Stratford Mountain Till of Koteff and Pessl, 1985). However, at this locality it shows an unusual degree of stratification in the form of silt-sand lenses. Perhaps it is a water-lain



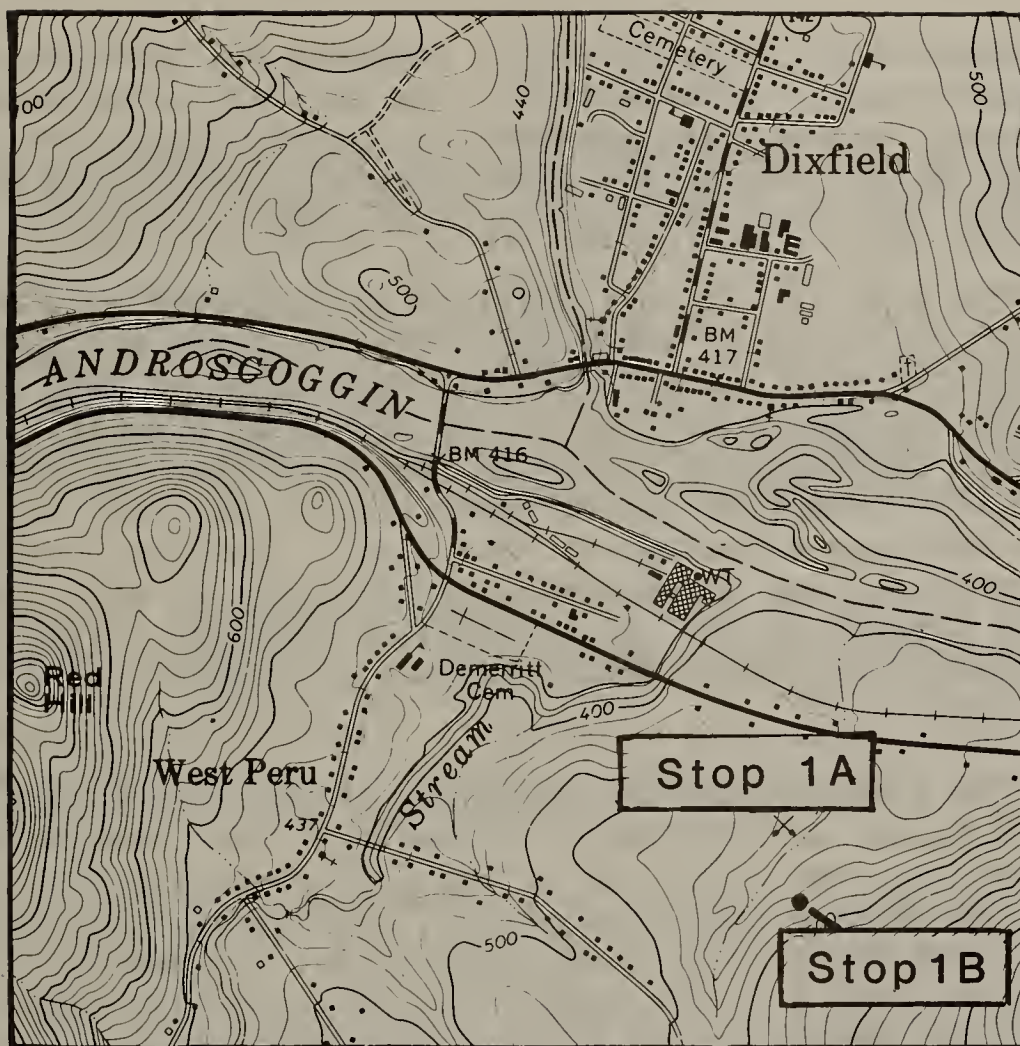


Figure 1. Location map for stops 1-A and 1-B

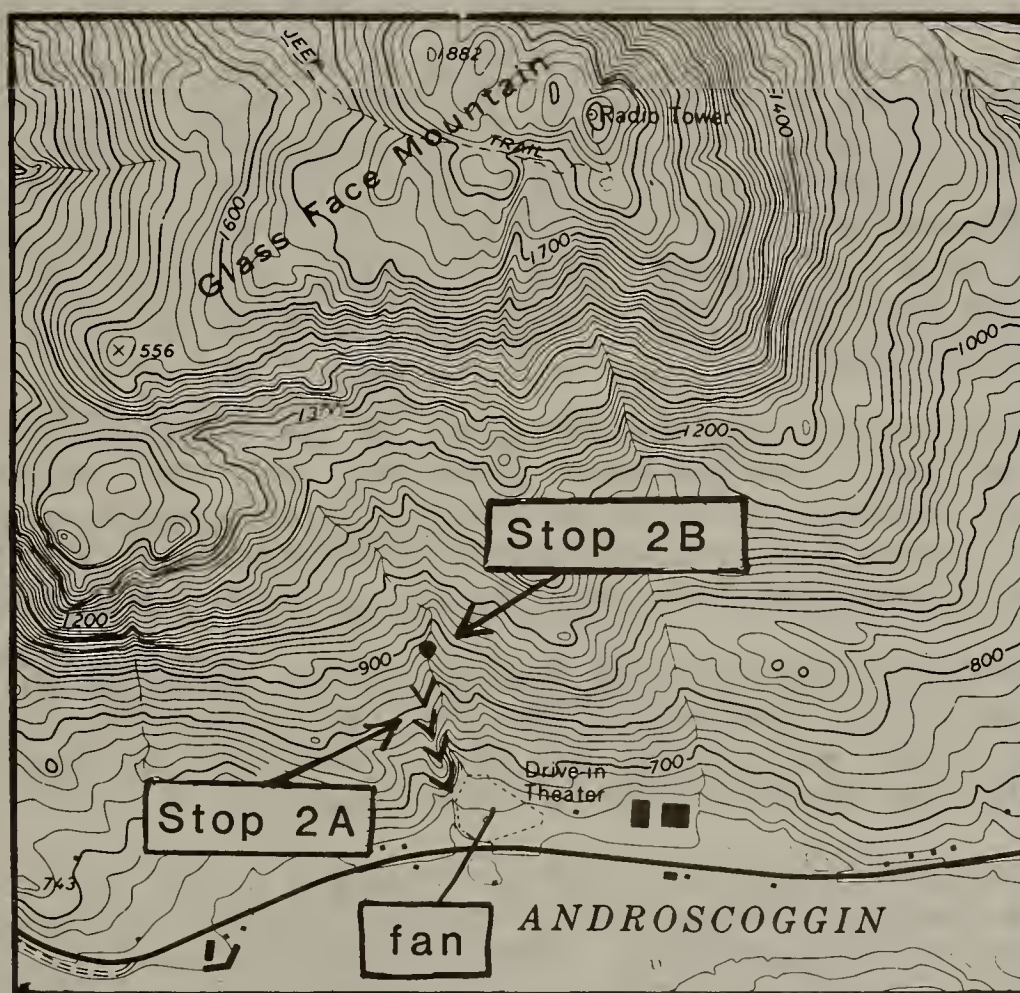


Figure 2. Location map for stops 2-A and 2-B. Chevrons indicate ice-channel filling.



deposit, since other units at Stop 1 indicate the presence of ponded water during deglaciation. Alternatively, the stratification suggests a possible basal melt-out origin. The presence of laminae draped over stones in the till support the latter theory, but more diagnostic structures must be found before the origin of this unit can be determined with confidence.

(2) The till is abruptly overlain by laminated silt and very fine sand. This unit probably can be correlated with the thick section of glaciolacustrine sand and silt located just south of here (Stop 1-B). It is believed to have been deposited on the bottom of a small glacial lake that was dammed in the Androscoggin Valley in the West Peru-Dixfield area. The outwash at 460+ ft that underlies the northern part of Dixfield village (Fig. 1) may be a delta that was deposited in this lake. The only possible outlet for the lake was directly down the Androscoggin Valley. It most likely was dammed by a temporary barrier of till or ice-contact sand and gravel deposits in the narrows between Whittemore Bluff and Morrill Ledges (2.5-3.0 km downstream from Dixfield).

(3) The glaciolacustrine sediments of Unit 2 are overlain by pebbly sand, which is not well exposed. Formerly it was observed to thicken to several meters at the north end of the pit, where much of the section has been removed. The surface elevation of this sand deposit is 420-440 ft; it correlates with the flat surface underlying the cemetery in West Peru. Fluvial cross-bedding in the sand dipped south to southeast in the north end of the pit, and pebble gravel with cross-bedding dipping east was noted in a railroad cut in the same unit just northeast of here.

Unit 3 is interpreted as glacial outwash or a postglacial stream terrace. In either case, it postdates the downcutting of the barrier that impounded the lake. Elevations of glaciofluvial deposits a short distance upriver (between Dixfield and Mexico) are in the 460-500 ft range, and probably are on-grade with the sand plain at West Peru. Therefore, Unit 3 is considered more likely to be glacial outwash. This kind of interpretive problem is common in narrow segments of the Androscoggin Valley, where Holocene erosion by the river has left only fragments of terraces. Given the 20-ft contour interval of the topographic maps and frequent lack of genetic criteria, it is difficult to determine the origin and age of some of the waterlaid units, or to correlate them along the valley.

Stop 1-B is the long washout gully in the logging road that leads across the brook and uphill to the south from Stop 1-A. This washout and the adjacent pit expose a total of at least 14 m of glaciolacustrine sediments. The deposit consists of well-stratified, subhorizontal sand and silt beds. The following characteristics were noted in these exposures:

-- Dark-colored finer-grained sets of thinly laminated silt and silty very fine sand beds occur throughout the section, indicating episodic low-energy conditions. These silty units are best exposed in the upper part of the gully, where they have a somewhat regular spacing of up to a few decimeters.

-- Ripple-drift lamination (current ripples) occur throughout the section in the sandy intervals. "Type A" and "Type B" ripples are present, as well as silt drapes (Ashley and others, 1985). Medium to very coarse sand seems to be more abundant higher in the section, suggesting higher-energy conditions with time.

-- The typical sedimentation pattern is interrupted at several levels in the section, where there are coarse channel fills and/or deformed and brecciated silt-sand beds. Midway up the gully, for example, laminated very fine sand with water-escape structures is disconformably overlain by a cross-bedded lens of coarse sand. The sand contains silt clasts and angular pebbles (the latter to 4.5 cm in diameter). This zone is overlain by a thin laminated silt-sand unit, which laterally becomes mixed and then truncated by subaqueous slumping, followed in turn by an overlying zone of intensely collapsed beds. The entire disturbed interval is about 1 m thick, and is both overlain and underlain by undeformed beds.

-- The directions of sediment transport indicated by ripples and cross-bedded channel fills vary greatly. Fifteen measurements taken randomly throughout the section ranged from  $20^{\circ}$  to  $350^{\circ}$ . Two-thirds of these measurements were in the northwest or northeast quadrants, while the remainder indicated south to southeastward transport. Additional measurements might reveal an even greater range of current directions.

Considering similarities to known glaciolacustrine deposits in New England, the sediments in this exposure are believed to have been deposited on the floor of a glacial lake. The sequences of rippled sand, beds/lenses of coarse to pebbly sand, and deformation structures indicate a nearby sediment source and dynamic environment. This site may have been located upon or close to a delta front. The remnant of an inferred delta is preserved across the river in Dixfield, as noted above. Since there are no dropstones in the lake sediments at Stop 1-B, the delta probably blocked the valley in front of the glacier margin and thus inhibited the development of icebergs.

The origin of the erosion features and coarse sand units, especially the fluvial-looking channel fills with pebbles and brecciated silt clasts, is uncertain. These features suggest the occurrence of sediment gravity flows from the nearby delta. The flows would have been triggered by slope failures on the unstable delta front. Perhaps dewatering of the lake-bottom sediments (indicated by water-escape structures) caused fluidization of beds in some cases. The frequency of northward sediment transport is another problem, since the alleged delta lies to the north of Stop 1-B, and there is only a till-mantled hillside to the south.

## STOP 2: Glass Face Mountain, Rumford

This is a two-part stop, located north of U. S. Route 2 on the lower slope of Glass Face Mountain. The large parking lot next to the highway is the former site of the drive-in theater shown on the Rumford Quadrangle (Fig. 2). This parking lot is situated on the surface of an alluvial fan deposited by the unnamed brook that flows down the hillside into the Androscoggin River.

Stop 2-A is the steep-sided ridge in the woods northwest of the parking lot (Fig. 2). Pit exposures show that the ridge is composed of gravel. Together with the morphology, this indicates that the deposit is an esker (or at least an ice-channel filling). There are numerous boulders on the ridge crest near the uphill end of the esker, as would be expected if meltwater flowed southward toward the valley and dropped the heaviest part of its load in the upper end of the ice tunnel.



This esker is one of several types of glacial features that typically occur on the distal (lee, or down-ice) sides of bedrock hills in the high-relief terrain of southwestern Maine. These features chiefly comprise meltwater channels (generally trending downhill, normal to contours) and various forms of hummocky moraine consisting of ablation till and/or sand and gravel. Less common are short, distinct ice-channel fillings like the one seen here. G. H. Stone described similar features, which he called "hillside kames" (Stone, 1890) and "hillside eskers or eskers" (Stone, 1899). He noted their abundance in the hills of western Maine, and their restricted distribution on south-facing slopes.

The questions raised by Stone concerning hillside eskers are relevant to the Glass Face Mountain deposit. He discussed the meltwater paths leading to these eskers, and the absence of similar features on proximal slopes. Stone concluded that they formed during a late stage of deglaciation, when "the ice north of the hills was still high enough to enable its drainage waters to flow southward over the hills" (Stone, 1899, p. 366). It is curious that the esker on Glass Face Mountain simply terminates against the hillside. It is slightly offset from the adjacent ravine, and there is no meltwater channel leading to its upper end. Glacial debris presumably was carried to this site through an englacial tunnel higher in the ice to the north or west. No continuation of the esker has been found down the valley. However, there is another ice-contact gravel deposit on the hillside just west of here, consisting of irregular hummocks that extend up to similar elevations (over 800 ft).

The genesis of lee-side channels and associated deposits needs more work in this part of Maine. These features may have been initiated subglacially in sites of reduced stress on the distal sides of hills as the ice was thinning. Accelerated melting and development of tunnel networks could have occurred in these locations as hills emerged and ice masses stagnated in valleys. However, the timing and formation of lee-side channels, eskers, and hummocky moraine deposits within this broad framework are not well understood.

At Stop 2-B, we will examine the ravine along the brook just east and northeast of the esker. This ravine is bedrock-floored along much of its length, and shows deep erosion of glacial sediments by a small, high-gradient intermittent stream during Holocene time. There are several waterfalls along the brook, including the one at Stop 2-B. The steeply dipping foliation of the metamorphic bedrock, striking transverse to the brook, provides structural control for this waterfall. Note that the upturned edges of the foliation layers have not been incised or greatly abraded, as we might expect if a debris-laden glacial stream had carved the ravine. On the other hand, many of the boulders along the brook can be moved on this steep slope during major floods (J. S. Kite, pers. comm.).

A small exposure of lodgement till occurs in the stream bank at Stop 2-B. The limited outcrop, and rusty mottling produced by near-surface weathering, preclude easy identification of this till as late Wisconsinan Stratford Mountain Till vs. the earlier Nash Stream Till (see Koteff and Pessl, 1985). The sheltered location in the lee of the mountain would be a likely site in which to find a remnant of pre-late Wisconsinan till.

When walking back to the highway, notice the narrow alluvial surface that occurs locally along the brook. It terminates at the alluvial fan next to Route 2. Many such fans have been deposited at the mouths of steep tributary streams in the upper Androscoggin River basin.

### STOP 3: Weston Pit, Rumford

The Weston Pit (Fig. 3) exposes up to 18 m of well stratified sand. Part of the pit face was freshly excavated in 1983. The material exposed at that time consisted of planar beds of fine to very coarse sand. The lower part of the section contained thin interbeds of pebble-cobble gravel and minor silt. "Starved ripples" have been noted, and boundaries between sets of beds are locally unconformable. The dip direction of bedding varies between east-northeast and south. No ice-contact deformation has been found here.

Internally, this deposit has the foreset bedding characteristics of a delta. It shows one or more foreset lobes built in an eastward (downvalley) direction. However, a delta topset unit is lacking, and the upper surface of the deposit slopes 8-10° eastward in conformity with the foresets. The Weston deposit is interpreted as a subaqueous glaciolacustrine fan, based on its close similarity to submarine fans of coastal Maine and lacustrine fans in other New England localities. The fan radiates from an apex located at the 780-800 ft col west of here, between the mountain to the north and a small bedrock knob to the south. A meltwater channel, with a boulder lag on its floor, traverses this col (Fig. 3).

Much sand also occurs around the south side of the rock knob, along the broad curve in Route 2. Though not apparent on the topographic map, a distinct terrace wraps around the hillside approximately 50 ft above the road, and there is possibly a higher terrace as well. These terraces resulted from the lowering of the lake in which the fan had been deposited. The original lake level is unknown, since deltas or shorelines have not been found. It presumably was impounded by glacial ice remaining in this narrow part of the Androscoggin Valley, and may have been a very local, ephemeral water body.

Coarse gravel is absent or rare in the section exposed at the Weston Pit. Most of the gravel derived from melting ice in this area was channeled into the esker system that comes down the Ellis River valley and crosses the Androscoggin about 2.5 km west of here. It is unclear why a large quantity of sand was diverted eastward to build the fan seen at this stop. The fan probably was deposited from the mouth of an ice tunnel where the glacier margin lay against the 780-800 ft col mentioned above.

### STOP 4: Red Hill Meltwater Channels, Rumford

Some of the most spectacular meltwater channels in this part of Maine are located on the west side of Red Hill (Fig. 4). The longest of these channels are plainly visible in the pastures belonging to the Kimball farm. They are up to 10 m or more in depth, and all terminate at the base of Red Hill. The channels are incised in thick bouldery till; no bedrock outcrops were seen on their floors. Unlike other types of meltwater channels, which slope obliquely across hillsides and follow the gradients of former ice margins, the Red Hill



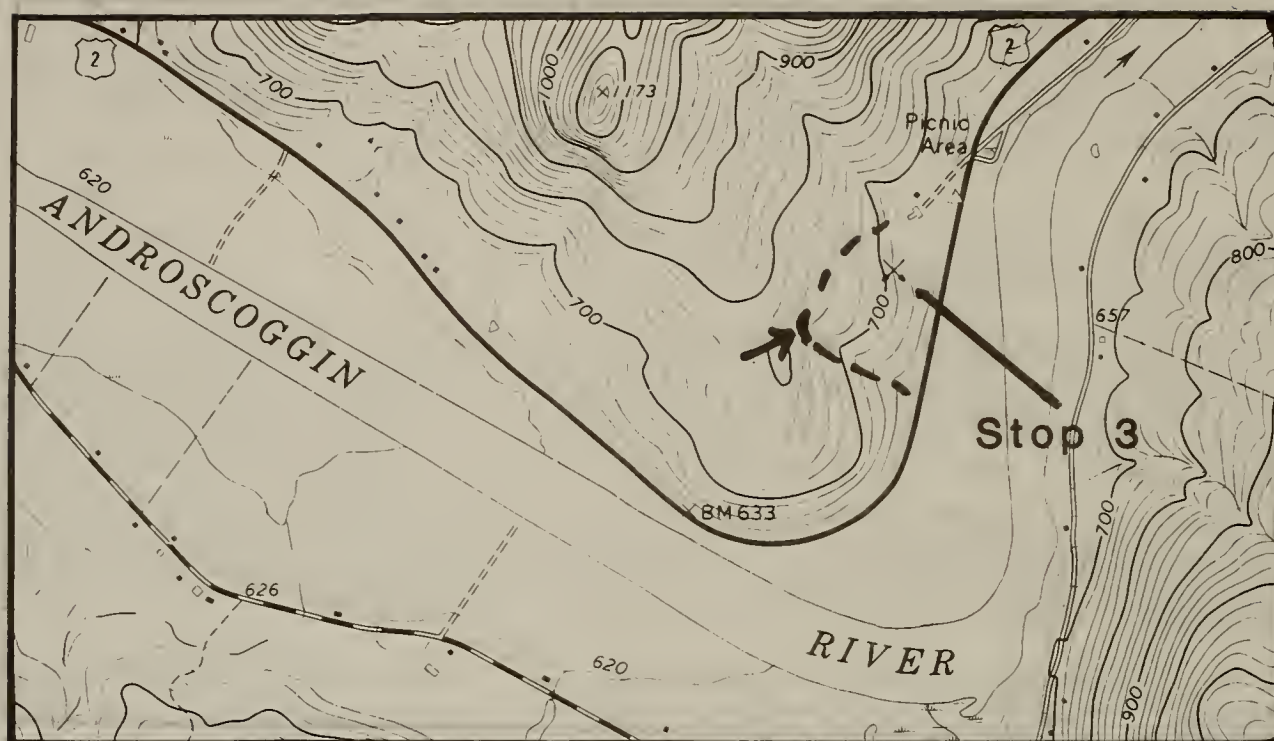


Figure 3. Location map for stop 3. Dashed line shows approximate boundary for glaciolacustrine fan seen at Weston Pit. Arrow indicates feeder channel.

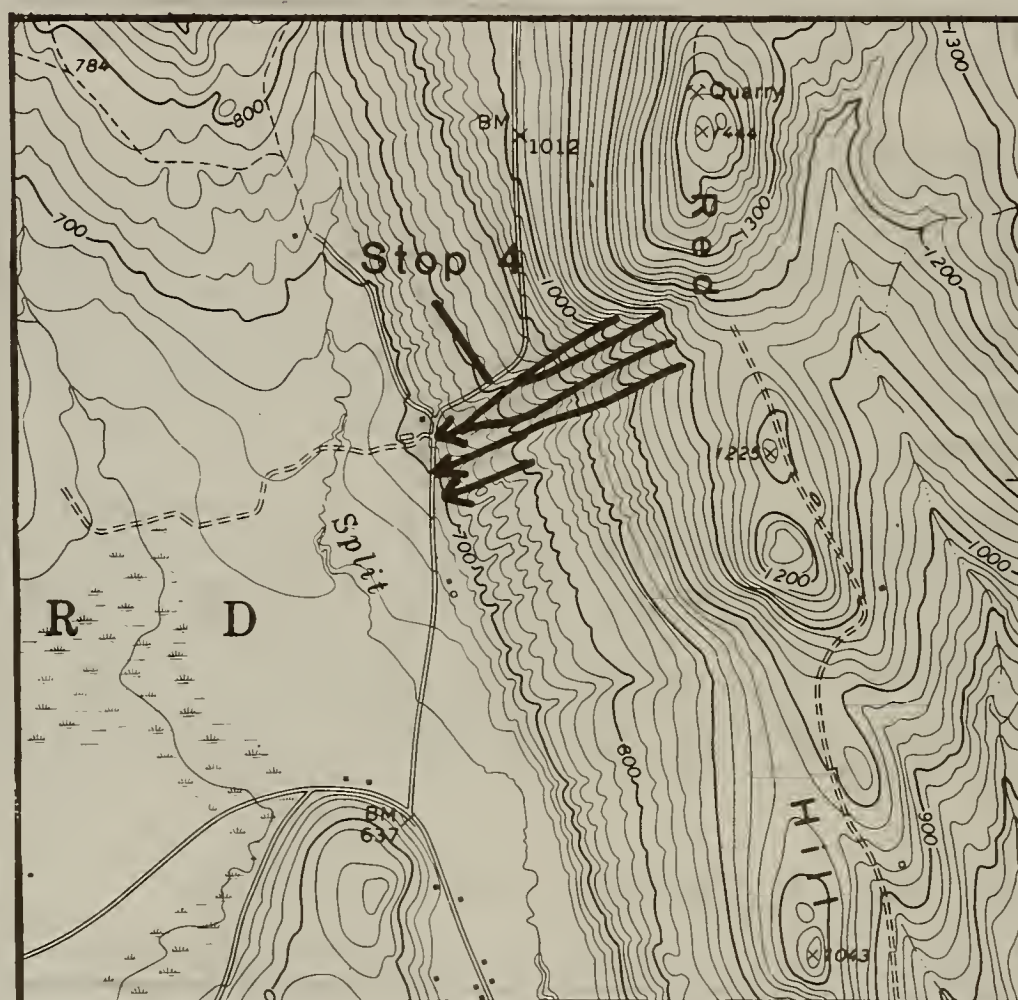


Figure 4. Location map for stop 4. Arrows indicate meltwater channels seen in open field. Other channels occur in woods to south.

channels plunge directly downhill, normal to the topographic contours. They are parallel and straight to slightly sinuous. In some places the divides remaining between channels are only a few meters across. Two channel junctions occur uphill from the Kimball farm, one of which is very near the bottom of the hill (Fig. 4). All observed junctions are accordant.

The upper ends of the central and southern channels become shallower and terminate before reaching the ridge crest. The northern channels extend farther uphill and end beneath a steep bedrock slope on the north peak of Red Hill. The full length of the ridge crest has not been explored, but observations to date have not revealed any channels that cross the top of the ridge.

The lower ends of the channels terminate at the eastern margin of a broad, gently sloping valley floor. This surface is underlain by Holocene alluvium deposited partly along a former course of the Ellis River, and partly by alluvial fans at the mouths of steep upland streams to the north. On air photos, it can be seen that small, poorly defined channels extend out onto this alluvial plain from the lower end of the Red Hill channels. There may be considerable runoff from the Red Hill channels during periods of heavy rainfall, during which sediment washes onto an alluvial fan west of Kimball Road. However, it is unlikely that postglacial erosion by meteoric streams has been sufficient to carve the deep, closely spaced channels seen on Red Hill.

Several questions concerning these channels remain to be answered:

- Assuming a glacial origin, what was the meltwater drainage path above and below the Red Hill channels?
- Were the channels cut subaerially or subglacially?
- Were they cut simultaneously or in sequence?
- Is their topographic setting in the lee of the north peak of Red Hill significant in terms of glacial drainage processes?

Since the central and southern channels do not straddle the ridge crest, it is likely that meltwater descended onto the hillside from a higher position in the glacier, thus requiring an ice cover on Red Hill. If so, we can infer that the channels formed subglacially. Meltwater drainage was focused on the lee side of the north peak when the northern channels were cut, but the absence of lateral channels on the hillside to the northwest suggests that meltwater continued to plunge down into the subglacial channels from higher within the glacier.

The systematic close spacing of the channels, and the relative timing of channel cutting, are problematic. Perhaps the pattern of subglacial tunnels was initiated along parallel crevasses in the ice? The subsequent path of the meltwater and debris exiting the channels is equally uncertain. No other glacial drainage indicators occur in the immediate vicinity, unless they are buried under the valley alluvium. It is possible that the meltwater streams connected with the esker system in the nearby Ellis River valley to the west.



## STOP 5: Chadbourne Pit, Newry

The Chadbourne Pit is located next to the confluence of the Bear River and Stony Brook with the Androscoggin. A complex deglacial history is recorded in this pit and the surrounding area shown in Figure 5. The reconstruction of this history is still in progress, but a tentative sequence of events is listed below. Some of these events probably overlapped in time.

(1) An esker system developed in the Bear River valley (along Route 26). It extends eastward down the Androscoggin Valley, though the discontinuous esker segments may not have been deposited simultaneously. The topographic map (Fig. 5) shows two segments in the area of the Chadbourne Pit. The ridge of the southeastern segment has been obliterated by expansion of the pit, while the segment to the northwest has recently been cleared of trees and shows excellent esker morphology (as of 1989).

Remnants of esker deposits comprise the oldest stratigraphic unit seen in the pit. An exposure of this unit in the northwest end of the pit consists of poorly sorted, non-stratified, locally openwork pebble-boulder gravel. Minor sand beds in this section are steeply collapsed. Other exposures of collapsed sand and gravel presently occur along the northeast wall of the pit. In the latter sections, the deformed unit extends up to 5 m above the pit floor. It is unconformably overlain and incised by fan gravel (discussed below), and probably is an eroded remnant of an esker.

(2) Ice-contact sand and gravel was deposited on the north side of the Androscoggin Valley. This episode is recorded by the high 740-780 ft terrace east of Stony Brook (Fig. 5). The terrace probably formed in contact with stagnant ice remaining on the south side of the valley. At least part of this deposit is deltaic. A pit on the Newry-Hanover town line (now a condominium development) exposed topset/foreset/bottomset beds in a delta graded to a lake level at about 740 ft. (location "d" on Figure 5). Lake-bottom sand and silt occurs in the gully just east of here.

A lacustrine unit that overlies esker sediments in the Chadbourne Pit may have been deposited in the same water body as the delta mentioned above. As seen in various sections during recent years, this unit consists of at least 4-5 m of well stratified sandy foreset beds, and sandy to silty, laminated lake-bottom sediments. The latter contain ripple-drift lamination, water-escape structures, and rare dropstones. Faulting locally occurs in the lacustrine unit, especially near the esker segments. Like the esker unit, the lacustrine sediments are disconformably overlain by fluvial gravel of the Stony Brook fan.

(3) Lowering of base level occurred as residual ice dissipated in the Androscoggin Valley. As this lowering was in progress, the channel at 720-740 ft on the southeast side of the valley (Fig. 5) probably was carved by meltwater from an ice mass to the south. Failure of the ice-dammed lake in the area of the Chadbourne Pit was followed by deposition of fluvial gravels at elevations in the 650-700 ft range. The channel cut in till southwest of Newry would have formed during this period -- see Figure 5).





Figure 5. Location map for stop 5. Symbols are as follows: arrows - meltwater drainage routes; chevrons - esker segments; d - ice-contact delta; f - outwash fan; O - outwash surface in Bear River valley.



The gravels were derived from two sources. One part is outwash deposited by meltwater flowing down the Bear River valley. The remainder comprises a large fan deposited at the mouth of the Stony Brook valley. A deep meltwater channel was cut in the latter valley. It can be traced uphill to a group of smaller channels, including a prominent one that cuts across the ridge north of Bear Mtn. (Fig. 5). An ice margin lay against the northeast side of this ridge and drained meltwater into the Stony Brook basin. However, meteoric streams in the steep headwaters of Stony Brook may have also contributed significant volumes of sediment ("inwash") to the fan. Regardless of the water source, much of the fan is believed to have been derived from erosion of thick till deposits that occur in the basin, rather than coming directly from the glacier margin.

The Stony Brook fan impinged against the high esker ridge northwest of the Chadbourne Pit, and wrapped around both ends of the esker segment. This accounts for the seemingly anomalous northwestward slope of the fan remnant that forms the highest terrace on the southwest flank of the esker, near Route 26. The western part of the fan merges with the Bear River outwash at about 780 ft. Walking northeastward up the Stony Brook fan, the gradient of the fan surface is evident. Test pits reveal sandy, angular pebble-boulder gravel.

In the Chadbourne Pit, the Stony Brook fan disconformably overlies both the esker and glaciolacustrine sediments described above. The contact displays considerable relief due to channel erosion at the base of the fan unit. This unit is presently well exposed along the northeast side of the pit. It consists of 1-3 m of gravel and pebbly sand. The fan sediments are variably well stratified, and show fluvial cross-bedding in cut-and-fill structures. In one part of the pit, the coarsest fan gravel (locally openwork pebble-cobble gravel) occurs in the bottoms of channels. Unlike the older sediments seen here, the fan deposit has not been disturbed by ice-contact collapse.

(4) Downcutting occurred in the valley when meltwater streams no longer supplied large volumes of sediment, and terraces formed in the Bear River valley at elevations between 640 and 660 ft. Two terrace levels can be distinguished along Rte. 26 in the fields southwest of the Chadbourne Pit. At least the higher, and perhaps both, of these terraces were cut by the postglacial river, though the gravelly sediments underlying them may be largely of glacial origin. A stream cut on the Bear River southwest of the pit shows an excellent exposure of this coarse, poorly sorted gravel, some of which is collapsed ice-contact material.

Downcutting also occurred on the Stony Brook fan, causing a subtle terracing of the fan surface. The part of the fan just west of Stony Brook appears to be graded to the same level as the higher Bear River terrace (about 650 ft).

(5) Beginning in late-glacial time, the Androscoggin River cut through a till barrier at the narrows just south of Newry (Fig. 5). This gap may have been the final spillway for glacial Lake Bethel (Thompson and Fowler, in press). Lake Bethel emptied as the spillway was eroded and the modern river drainage was established. The present flood plain of the Bear and Androscoggin Rivers is at 620-640 ft near the junction of Routes 2 and 26.

#### STOP 6: Screw Auger Falls, Grafton

The scenic gorge at Screw Auger Falls is up to 13 m deep, and has been cut into granite and granite pegmatite. It is doubtful that the present Bear River could have carved this gorge. It is interpreted to have been cut by a debris-laden glacial stream that was "superimposed through dead ice onto the outcrop" (Brewer, 1978). This process probably occurred subglacially, with the meltwater under great hydrostatic pressure. The same model explains the development of other gorges farther upstream at Mother Walker Falls and Moose Cave.

#### STOP 7: Grafton Notch Meltwater Channel and Deposits, Grafton

Grafton Notch is one of several deep notches that penetrate the Mahoosuc Range. These notches, and numerous other lesser gaps in the hills of western Maine, played important roles in conducting glacial meltwater during recession of the late Wisconsinan ice sheet (Thompson and Fowler, in press). There are three ways in which they functioned, though not all of these events occurred at every locality:

-- Meltwater at first could flow through the notches subglacially. A through-going esker system is the best evidence that this actually happened. Some eskers are continuous through gaps, while others are interrupted by them.

-- As the glacier withdrew through each notch, there would have been a very brief period when the ice margin stood in the notch and outwash was transported downvalley to the south. This phase is rarely evident in the notch itself (unless a moraine is preserved to mark the ice-margin position), since stream gradients immediately below the notches were generally too steep for outwash deposition. Any related deposits usually are located farther downstream.

-- Continued ice retreat resulted in short-lived glacial lakes being dammed between the glacier margin and the mountains to the southeast. The notches then served as spillways for the lakes. Ice-contact deltas at the same elevations as the notches, and sometimes immediately behind them, are the surviving indication of this phase of deglaciation. Although not as numerous or large as the glacial-lake deltas in ponded drainages of southern New England, the deltas and other features in this region of Maine provide consistent evidence for progressive northwestward recession of the ice margin (Thompson and Fowler, in press).

Stop 7 is located on the drainage divide in Grafton Notch (Fig. 6). The woods road leading northwest from Route 26 quickly crosses the upper end of a meltwater channel (now a swampy area) and then follows the west side of it. This channel can be traced southward for about 0.7 km. It is not deeply incised, and may have carried meltwater for only a brief time. At the north end of the channel, and bisected by a brook that crosses the woods road, there is a small, flat-topped, kettled sand and gravel deposit. This feature is tentatively identified as an ice-contact delta ("d" on Fig. 6). It is graded to the head of the channel, and connects with an esker ridge to the north. A



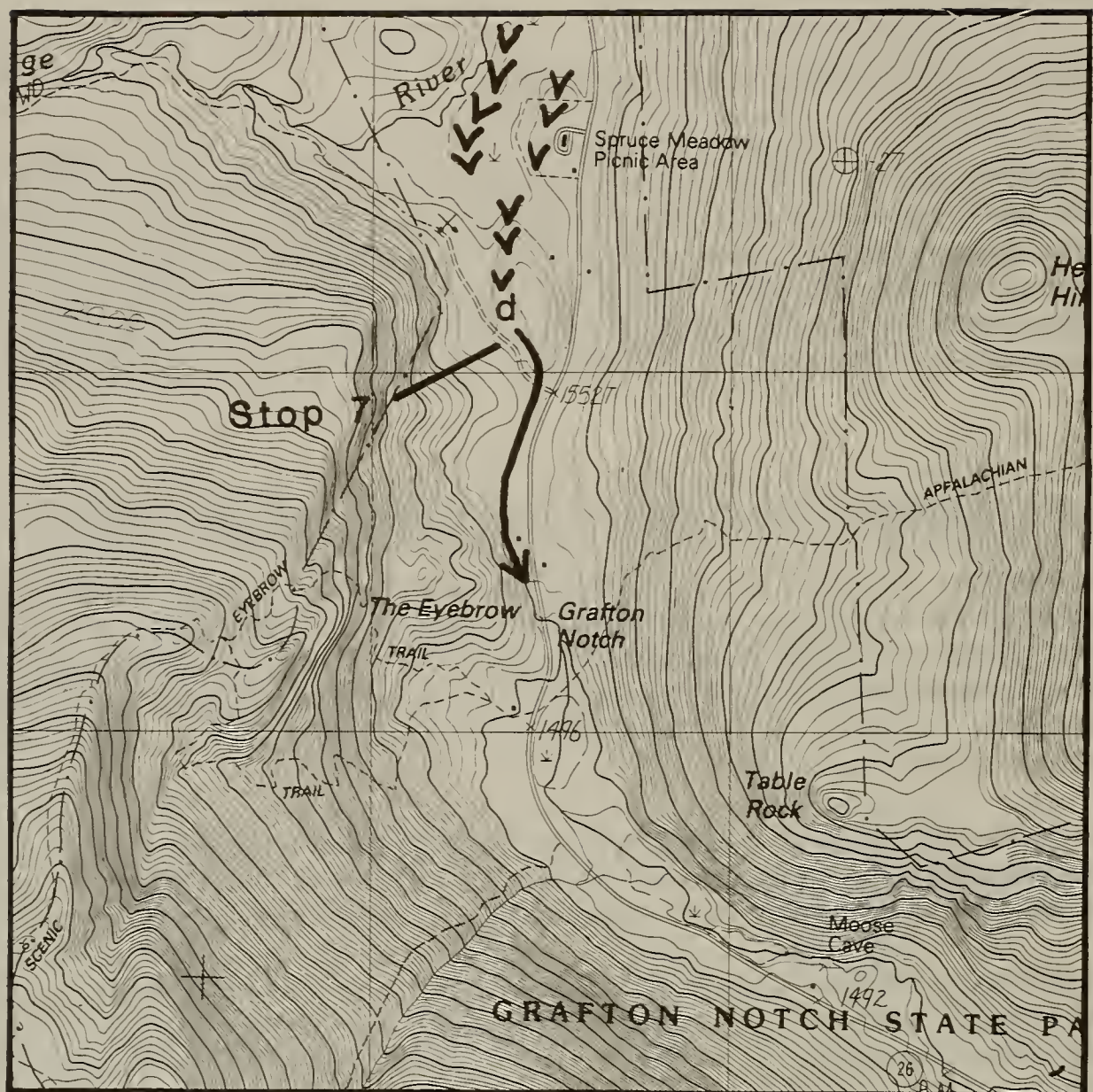


Figure 6. Location map for stop 7. Symbols are as follows:  
 arrow - meltwater channel; chevrons - esker segments,  
 d - ice contact delta (?).

bedrock knob protrudes above the floor of Grafton Notch on the east side of the delta. The esker that fed the delta is part of a multi-segment esker system that leads southward to the notch from the Swift Cambridge River valley. A bouldery ridge -- possibly a moraine -- begins in the woods across the road from the delta and extends uphill to the southwest.

The Grafton Notch channel was the highest and earliest spillway for glacial Lake Cambridge. This lake was named by Leavitt and Perkins (1935). Glacial recession from the proximal side of the Mahoosuc Range uncovered successively lower outlets for Lake Cambridge until it finally emptied into the upper end of the Androscoggin Valley in New Hampshire (Thompson and Fowler, in press).

#### ACKNOWLEDGMENTS

The Maine Geological Survey provided logistical support throughout this study. The author is grateful to Craig Neil and Julie Poitras, Maine Geological Survey, for assistance in preparing the location maps accompanying this paper. The Chadbourne Lumber Company, Kimball and Weston families in Rumford, and other landowners were very helpful in allowing field trip stops on their property.

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# ITINERARY

Assembly point: South side of Route 108, next to cemetery in village of West Peru. The itinerary is covered by Maps 18, 19, and 10 in the Maine Atlas. Topographic map coverage for the field trip stops is provided by the Dixfield, Rumford, East Andover, Bryant Pond, Bethel, Puzzle Mountain, and Old Speck Mountain 7.5-minute quadrangles.

Total Mileage	Mileage from Previous Point	
0.00	0.00	Four-way jct. in West Peru village. Go east on Rte. 108.
0.55	0.55	Turn R onto dirt road and drive 0.1 mi south into sand pit.
		<u>STOP 1: West Peru Sections (Dixfield Quadrangle)</u>
		<u>STOP 1-A:</u> Till exposure in southwest corner of pit area. <u>STOP 1-B:</u> From south end of large pit, walk south on logging road, cross brook, and proceed to smaller sand pit and long gully resulting from erosion of road bed on hillside.
0.75	0.20	Return to Rte. 108. Turn L and drive west to Rumford.
6.30	5.55	Cross Androscoggin River at Rumford. Merge with U.S. Rte. 2 just past bridge, and continue straight (W) on Rte. 2.
10.45	4.15	Park on R shoulder of Rte. 2.
		<u>STOP 2: Glass Face Mountain (Rumford Quadrangle)</u>
		<u>STOP 2-A:</u> Walk north to back side of parking lot, then cross to west side of small brook and walk steeply uphill onto ridge crest (ice-channel filling) in woods.
		<u>STOP 2-B:</u> Walk up brook a short distance past upper end of ice-channel filling, until reaching waterfall and small till exposure.
		Continue west on Rte. 2.
13.90	3.45	Park on R shoulder of Rte. 2, and walk into sand pit on north side of road.
		<u>STOP 3: Weston Pit (Bryant Pond Quadrangle)</u>
		Continue west on Rte. 2.
16.25	2.35	Just past village of Rumford Point, note outstanding point bars to L, on opposite side of sharp bend in Androscoggin River.



16.40	0.15	Turn R (N) onto Jed Martin Rd. (sign obscured by bushes).
17.90	1.50	Note Meadow Bk. valley to west of road -- former course of Ellis River?
19.00	1.10	Turn R (E) onto Andover Rd. (no sign).
19.15	0.15	Turn L (N) onto Kimball Rd., and proceed north to Kimball Farm.
19.65	0.50	Park on R. Walk uphill through gate, into pasture on hillside. <u>Be sure to close gate; and do not use shovels in cow pasture!</u>

STOP 4: Red Hill Channels (East Andover Quadrangle)

Turn around at Kimball Farm and return by same roads to Rte. 2.

23.05	3.40	Rejoin Rte. 2. Turn R and continue west.
29.50	6.45	Just past Newry town line, turn R onto gravel road and drive north into large pit.

STOP 5: Chadbourne Pit (Bethel Quadrangle)

29.90	0.40	(Mileages will vary after driving around in the Chadbourne Pit.) Enter Rte. 26 from west side of pit. Turn R and drive north on Rte. 26.
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39.50	9.60	Turn L into State parking lot for Screw Auger Falls.
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STOP 6: Screw Auger Falls (Old Speck Mtn. Quadrangle)  
**No hammers or shovels at Stops 6 and 7!**

39.60	0.10	Return to highway. Continue north on Rte. 26.
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42.85	3.25	Park on R shoulder of Rte. 26, as far off road as possible. Walk west on woods road about 430 paces to ice-contact deposits at head of meltwater channel (in woods to R -- see location map).
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STOP 7: Grafton Notch (Old Speck Mtn. Quadrangle)

END OF TRIP

# TURBIDITES AND MELANGES OF THE MADRID FORMATION, CENTRAL MAINE<sup>1</sup>

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## INTRODUCTION

This field trip guide describes six of the most informative exposures of the Madrid Formation in central Maine (Figure 1). The Madrid Formation is an important key to the evolution of the Maine Appalachians, because it represents an influx of clastics from a new source shortly before deep marine sedimentation was terminated by the Acadian Orogeny. In addition, the Madrid and overlying Carrabassett Formations include abundant melanges and olistostromes; these rock types either were not recognized or not emphasized by previous workers, but have an important bearing on the regional tectonic history.

Figure 2 illustrates the stratigraphic position and across-strike facies relations of the Madrid Formation in western Maine, according to Hatch and others (1983). This widely accepted cross section implies that sedimentation took place in a relatively quiescent tectonic setting not complicated by plate boundaries, leading to a relatively simple stratigraphy. Whereas Figure 2 was based largely on regional lithofacies relations of complexly deformed, fossil-poor metasedimentary rocks, our complementary studies have focused on detailed analysis of sedimentary facies and paleocurrents. We question some implications of Figure 2 and suggest that at least the upper Madrid and Carrabassett Formations were deposited in a convergent tectonic setting, that the stratigraphy is complex, and that lithologic contacts are probably diachronous.

The present study is one aspect of an ongoing investigation of the Siluro-Devonian depositional history and tectonics of the Maine Appalachians (Bradley 1983, 1987; Hanson 1983, 1988; Hanson and Bradley 1989; the latter paper ). Research was funded by NSF Grant EAR-88-03233. The USGS Branch of Alaskan Geology provided Bradley with official time for field work and the NEIGC trip; Hanson received release time from Salem State College for field work during the academic year. Critical reviews by S. M. Karl and M. L. Miller significantly improved the manuscript.

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<sup>1</sup> This manuscript has not been reviewed for conformity with U.S. Geological Survey stratigraphic nomenclature and ages.



## Bradley and Hanson: Madrid Formation

Figure 1. Map of central Maine showing the field trip route (dashed line with arrows and route numbers) and stop locations. The distribution of the main body of the Madrid Formation (fine stipple) and eastern facies of the Madrid Formation (coarse stipple; formerly Fall Brook Formation) is from Osberg and others (1985).

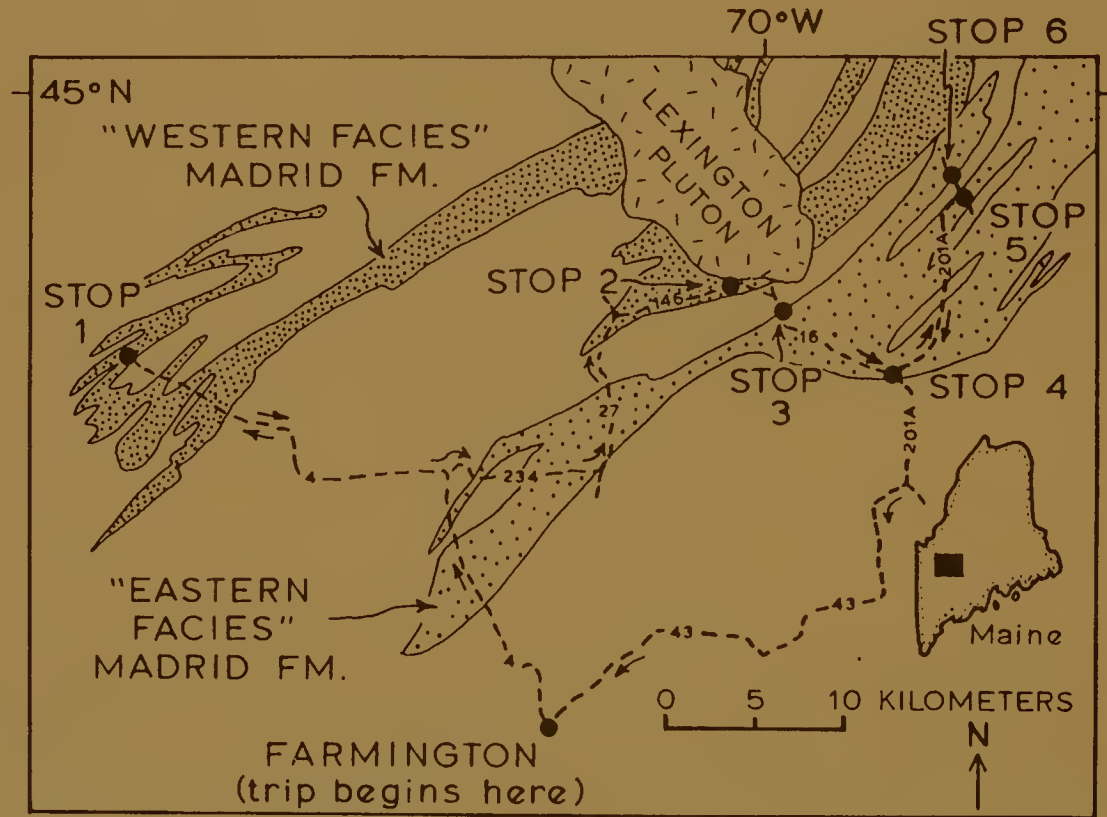
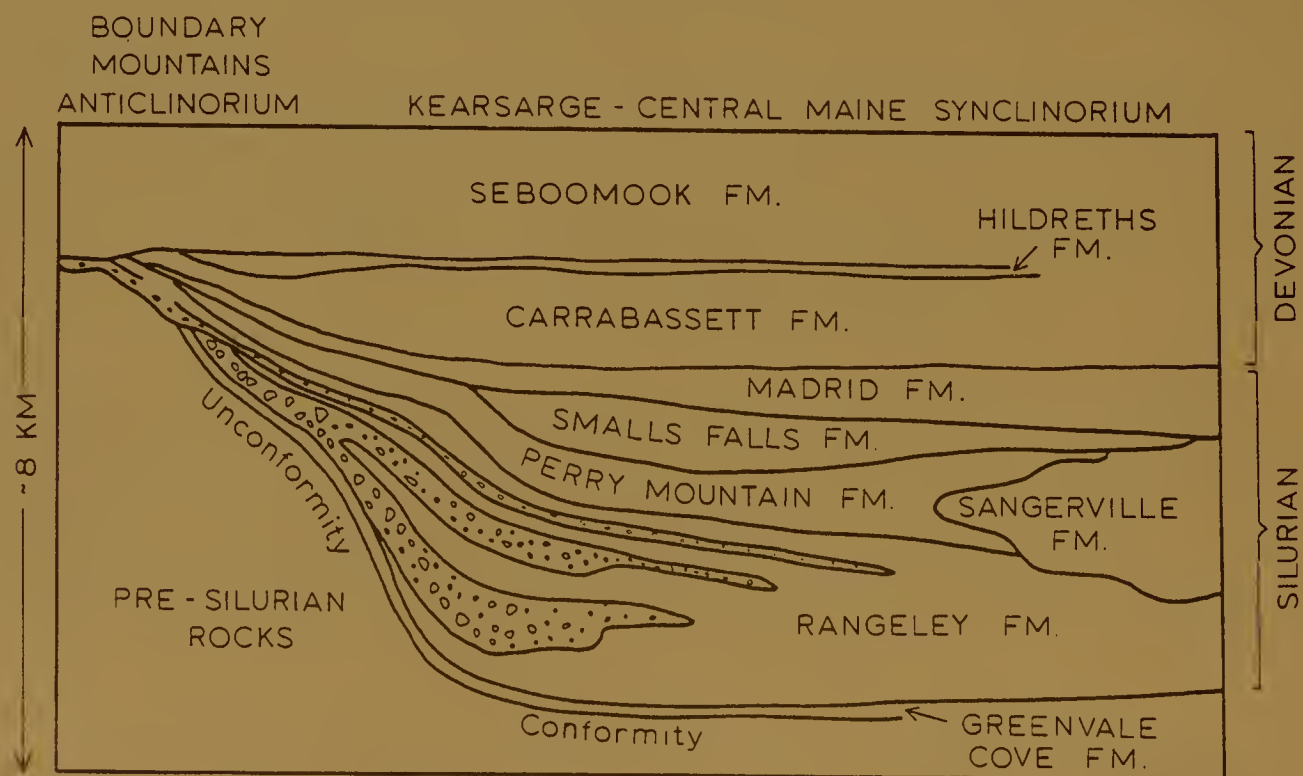


Figure 2. Stratigraphic cross section through the northwestern margin of the Kearsarge-Central Maine Synclinorium in the Rangeley area, as interpreted by Hatchand others (1983).



## GEOLOGIC SETTING

The Kearsarge-Central Maine Synclinorium is a 500 km-long, 100 km-wide belt of deep-water, Upper Ordovician to Lower Devonian strata that were intensely tectonized during the Devonian Acadian Orogeny. At least 8 km of strata (Figure 2) were deposited on basement of unknown type, in water depths below storm base and mostly below the CCD. Much of the present controversy over Acadian tectonics centers on whether or not the Kearsarge-Central Maine Synclinorium was the site of a pre-Acadian ocean basin, which closed by subduction during Silurian. The Acadian Orogeny resulted in extreme horizontal shortening, and it coincided with the end of long-lived volcanism in two linear belts that flank and parallel the Kearsarge-Central Maine Synclinorium. These first-order observations are the basis for interpreting the Acadian Orogeny as the result of arc collision following subduction of an oceanic Kearsarge-Central Maine Synclinorium, by one or another plate geometry (Bradley 1983). The Madrid and Carrabassett Formations comprise the youngest widespread units in the synclinorium<sup>2</sup> to be deposited before sedimentation gave way to contractional tectonic deformation, regional metamorphism, crustal thickening, anatectic melting, and uplift. This suggests that the Madrid and Carrabassett Formations were deposited in a convergent tectonic setting.

An Acadian-derived clastic wedge spread across eastern North America during Devonian times (Ettensohn 1985; Bradley 1987). Flysch and/or molasse units in this diachronous succession include the Temiscouata Formation and Gaspé Sandstone of Quebec, the Seboomook Group of northwestern Maine, the Littleton Formation of New Hampshire, and the Catskill "delta" of the central Appalachians. In all but the Catskill "delta" (which lay cratonward of the Acadian deformation front), flysch or molasse sedimentation was followed immediately by Acadian contractional deformation. Based on paleocurrent data and regional facies relations, we suggest here that the upper Madrid Formation represents the oldest manifestation of the Acadian clastic wedge, and that clastic progradation was already underway by Late Silurian.

Within the field trip area, upright, tight to isoclinal folds of Upper Ordovician to Lower Devonian strata dominate the map-scale structure. These folds are attributed to the main phase of the Acadian Orogeny and are associated with an axial planar foliation. Acadian regional metamorphism reached staurolite grade in the western part of the field trip area (Stop 1), and biotite grade in the eastern part of the field trip area (Stops 2-6). Acadian folds and metamorphic isograds are cut by numerous Acadian plutons including the 400 Ma. Lexington batholith (Figure 1; see Stop 2 for elaboration); hence the Acadian Orogeny was an Early Devonian event in the field trip area.

## STRATIGRAPHY

The Madrid Formation was formally named by Osberg, Moench, and Warner (1968). In the type area in western interior Maine, it consists of two members. The lower member, about 100 m thick, consists of "thin bedded calcareous metasandstone, metashale, and subordinate calc-silicate granulite" (Osberg and others, 1968, p. 251). The upper

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<sup>2</sup> Although the Hildreths and Seboomook Formations are shown in Figure 2 as regionally extensive, they are absent in much of the field trip area. The name "Kearsarge-Central Maine Synclinorium" is used here without endorsing the implication that the regional structure resembles a complex syncline.



## Bradley and Hanson: Madrid Formation

member, which is the focus of this field trip, is about 200 m thick and consists of "thick-bedded metasandstone and subordinate metashale" (Osberg and others, 1968, p. 251).

The Madrid Formation and its presumed correlatives have been traced from the type area (Stop 1) along most of the 500-km length of the Kearsarge-Central Maine Synclinorium. To the northeast at Stops 2-6, sedimentary structures are better preserved than at the type area. Some feldspathic, locally calcareous, turbiditic sandstones in the field trip area were originally assigned to the Fall Brook Formation during 15' quadrangle mapping (Pankiwskyj and others 1976; Pankiwskyj 1979) but were reassigned to the informal "eastern facies" of the Madrid Formation by Moench and Pankiwskyj (1988) to simplify the stratigraphic nomenclature. The eastern facies is lithologically similar to the upper member of the type Madrid; the lower member of the type Madrid Formation is not recognized in the east. Madrid Formation sandstones thicken from west to east (200 meters in the upper member in the type area), 600-1000 meters in the field trip area (Pankiwskyj and others 1976), and about 1500 meters in the Skowhegan quadrangle (formerly Brighton Formation; Ludman 1976).

Although the Madrid Formation is devoid of fossils, indirect correlations suggest a Late Silurian to Early Devonian age (Moench and Pankiwskyj 1988; Osberg and others 1985).

### TURBIDITE AND ASSOCIATED FACIES

The field trip will show that the eastern facies and the upper member of the type Madrid Formation mainly consist of siliciclastic turbidites deposited in a submarine fan environment. Individual beds consist of partial to complete Bouma sequences (denoted Ta to Te after Bouma 1962); sets of turbidite beds are grouped herein into facies according to a scheme adapted from Mutti and Ricci-Lucchi (1978; see also Hanson and Bradley 1989). Three main sandstone-rich turbidite facies are present: Facies B, C, and D. *Facies B* in the Madrid Formation consists of single or amalgamated sandstones that cannot be readily described in terms of Bouma sequences, and are typified by lateral thickness changes. The base of a typical bed is rapidly graded and contains sets of low-angle, scoop-shaped backset laminae, probably produced by antidune migration. The upper portions of most *Facies B* sandstones are apparently homogeneous and structureless; crude parallel laminae are locally present. Scarcity of clearly visible sedimentary structures may result, in part, from originally well-sorted sediment; if sediment is entirely homogenous, no primary structures will be evident regardless of the mode of transportation and deposition. *Facies C* in the Madrid Formation consists of classic turbidites, and includes complete (Ta-e) and top-missing (Tab, Ta-c) Bouma sequences. *Facies D* in the Madrid Formation consists of base-missing turbidite sequences (Tb-e, Tc-e, Tde). Beds are typically plane-parallel with little variation in thickness along strike.

In addition to turbidite facies, the upper Madrid Formation also contains slump deposits (*Facies F* of Mutti and Ricci-Lucchi 1978). Slumps in the Madrid Formation are characterized by contractional soft-sediment deformation, suggesting accumulation near the base of a submarine slope. The overlying, pelite-dominated Carrabassett Formation is also characterized by soft-sediment deformation (Hanson and Bradley 1989), but the style was largely extensional, suggesting deposition on a submarine slope. Slump deposits are recognized and discriminated from disrupted facies of tectonic origin by (1) chaotic, disharmonic folding; (2) stratabound geometry with enclosing, intact beds younging in a consistent direction both above and below the slump horizon; (3) a welded upper sedimentary contact; and (4) a detachment rather than a depositional contact at the base.



## LITHOLOGY AND PROVENANCE

Turbidites of Madrid Formation consist of locally calcareous quartz-plagioclase metasandstone and subordinate metapelite. Pankiwskyj (1979) reported mean Q:F:L ratios of 70:15:3 based on point counts of 7 sandstone thin sections; detailed petrographic studies are in progress. Considering composition only, the feldspathic sandstones of the Madrid Formation could plausibly have been derived from either outboard (Avalonia), inboard (Boundary Mountains Anticlinorium), along strike to the north (Aroostook County or New Brunswick), or along strike to the south (New Hampshire). In other words, the Kearsarge-Central Maine Synclinorium is flanked on all sides by possible source terranes; paleocurrent data appear to be more useful than petrographic data in establishing sediment dispersal patterns.

Many sandstones in the upper member and eastern facies of the Madrid Formation are calcareous, a fact that has proven invaluable in regional geologic mapping. Calcareous lithologies form both football-shaped concretions (within siliciclastic metasandstone beds) and beds of calcareous metasandstone; there is a gradation between discrete ellipsoidal concretions and entire beds. Pure metalimestones are absent; associated metapelites are noncalcareous. Throughout the field trip area, the calcareous concretions and beds have been metamorphosed to calc-silicate metamorphic assemblages, typically biotite+actinolite+clinozoisite+grossularite+diopside (Pankiwskyj 1979, p. 20).

The calcareous component is not yet adequately understood. We believe that the carbonate was originally detrital and was probably transported to the depositional site by turbidity currents. This is difficult to establish in the field trip area, where calc-silicate metamorphic minerals are commonly many times larger than the siliciclastic detritus, and primary sedimentary textures are lost. However, in chlorite-grade Madrid sandstones northeast of the field trip area (Spectacle Pond near Monson), carbonate grains appear to have been detrital and transported by traction currents, on the basis of carbonate concentrations in troughs of ripple-drift cross laminae in Bouma C divisions. At Stop 6, carbonate grains are concentrated in the graded bases of some coarse-grained sandstone beds; a detrital origin is likely. Detrital carbonate might either have originated by erosion of carbonate bedrock units, or in a shallow marine carbonate factory flanking the Kearsarge-Central Maine Synclinorium basin, perhaps along strike to the northeast (see paleocurrent discussion). The calc-silicate concretions are early diagenetic features that clearly predate folding; similar calcareous concretions of diagenetic origin are common in deep-water flysch sequences (e.g., the Valdez Group, a Cretaceous trench-fill deposit in Alaska; Nilsen and Zuffa 1982), and do not constitute evidence for a shallow-water origin for the upper Madrid Formation.

## PALEOCURRENT DATA

Previous workers (e.g., Ludman 1976) speculated that sedimentary filling of the Kearsarge-Central Maine Synclinorium occurred in two stages. According to this model, the older units were derived from the inboard side of the basin (e.g., northwest-derived conglomerates of the Rangeley Formation), and at least the youngest Devonian strata were derived from outboard. We are testing and refining this model through a regional paleocurrent and facies survey, with particular emphasis on the purported transition between inboard and outboard derivation.

Our paleocurrent data from the Madrid and Carrabassett Formations suggests that an outboard sediment source probably did exist during Late Silurian and Early Devonian time. Unequivocal evidence for outboard derivation is provided by paleocurrent data from



the Carrabassett Formation northeast of the field trip area (Hanson and Bradley 1989); sediment transport in the Carrabassett was mainly toward the northwest (across strike), with secondary modes to the northeast and southwest (along strike). At Stops 4, 5, and 6 in the field trip area, the dominant paleoflow direction in the eastern facies of the Madrid Formation was toward the southwest, parallel to tectonic strike. Paleoflow directions were obtained from cross-laminated Facies D turbidites, using standard structural corrections (single-tilt and double-tilt for non-plunging and plunging folds, respectively). Only generalized directions are quoted here because our results may be subject to revision by up to a few tens of degrees, pending more sophisticated analysis (strain correction and incremental plunge removal) now in progress.

### MELANGES

Melanges are far more common in the Kearsarge-Central Maine Synclinorium than the older literature suggests. The term *melange* is used here in the descriptive sense of Cowan (1985, p. 452) for a rock body composed of "fragments enveloped by a finer-grained matrix of mudstone". The Madrid melanges consist of partially to thoroughly disrupted metasandstones in a metapelite matrix and hence can be assigned to Cowan's (1985) Type I (in which exotic clasts are absent). The foliated matrix wraps around clasts.

Discrimination between a sedimentary or tectonic origin for the Madrid Formation melanges is hindered by the effects of Acadian regional folding with associated axial planar cleavage, and associated metamorphism. At two field trip stops, evidence for a tectonic origin can be seen from a progression from relatively intact beds, to beds cut by web structure, to a block-in-matrix melange containing turbidite bed fragments (Figure 3). The term *web structure* (Byrne 1984) is applied to an intricate network of cataclastic microfaults that pervasively cut individual sandstone beds; microfaulting results in layer-parallel extension and boudinage. Web structure is typical of tectonized turbidites in modern and ancient accretionary prisms (e.g., Lundberg and Moore 1986; Knipe 1986). Other features suggesting a tectonic origin include (1) mesoscale faults and high strain zones that parallel fragment foliation in the melange zone, and (2) an absence of features enumerated above that would be diagnostic of slump deposits.

Figure 3. Schematic diagram showing progressive fragmentation of an intact turbiditic sandstone bed, into a bed cut by web structure, to a block-in-matrix melange (Type I melange of Cowan 1985). Arrows show stratigraphic way up in beds and bed fragments.

a. Intact bedding



b. Web structure



c. Type I melange



## Bradley and Hanson: Madrid Formation

### SUMMARY

The upper Madrid Formation and the laterally equivalent eastern facies constitute an eastward-thickening, SW-prograding submarine fan complex of Late Silurian to Early Devonian age. The Madrid is overlain by olistostrome-dominated slope deposits of the Carrabassett Formation. Madrid paleocurrents indicate that sediment transport was dominantly toward the southwest, parallel to tectonic strike; Carrabassett paleocurrents record transport mainly toward North America. These observations, coupled with regional relations discussed earlier, suggest that the Madrid Formation was deposited in a migrating Acadian foredeep. We visualize an elongate, strike-parallel Madrid depositional basin that was flanked on the southeast by an northwest-advancing, submarine accretionary prism, atop which the Carrabassett Formation was deposited (Hanson and Bradley 1989). Within the KCMS, the area of Carrabassett sedimentation and deformation migrated to the northwest across the area of Madrid sedimentation, displacing facies belts laterally with time. Regionally, the locus of foredeep sedimentation migrated across- and along-strike, beginning at outboard and northeasterly locations, and ending at inboard and southwesterly locations. By the time the Catskill "delta" began prograding across the foreland, the Kearsarge-Central Maine Synclinorium had long since been destroyed and deformed. Within this context, the lower Madrid and Smalls Falls Formations (Figure 2) could represent the transition from inboard to outboard derivation; detailed studies of these key units are clearly warranted.

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## Bradley and Hanson: Madrid Formation

### ROAD LOG

The field trip will visit 6 localities. Stop 3 is optional, depending on time, weather, and interest. Please refrain from hammering unless sampling for research. Lunch will be at Stop 4B, where provisions can be bought.

Meet in front of the Geology Building at University of Maine in Farmington. Drive north about one-quarter mile on Route 4 to its junction with Route 143. Begin mileage here.

To keep all vehicles together, please wait for the car behind you.

Miles

0.0 Drive north on Route 4.

24.5 Madrid Village. Turn right over a narrow bridge and park. The outcrops of interest are exposed on private property along the Sandy River and its tributary, Saddleback Stream. Please check in at the house nearest the outcrops to ask for permission to look at the outcrops.

#### STOP 1 - Madrid Village

These outcrops at the type locality expose the transition from thinly-bedded, impure metalimestones and metapelites of the lower member of the Madrid Formation (along Saddleback Stream) to thicker-bedded, massive sandstones of the upper member of the Madrid Formation (along Sandy River). The section along Sandy River (Figure 4; this is the first published section from type locality) is mainly composed of amalgamated sandstone beds (Facies B). Convincing upward-thickening or upward-thinning trends are lacking. The absence of deeply channeled bases suggest deposition on a smooth-floored outer fan, while the predominance of amalgamated sandstones suggests deposition not far from the outlet of a feeder channel. The thick-bedded sandstones at first seem entirely massive and featureless, but on close inspection many beds display grading, rip-up clasts, convolute laminae, and large-scale, low-angle cross-laminae (backset laminae). The sandstones are typically very fine to fine grained; medium-grained sandstones are rare and restricted to the lowest few centimeters of sandstone beds.

24.5 Retrace route on Route 4 (S), past Phillips, to Strong.

38.5 Turn left (N) onto Route 145, cross the Sandy River.

38.75 Turn right (S) onto Route 149.

39.0 Turn left (E) onto Route 234; drive to New Vineyard.

44.2 Turn left (N) onto Route 27; drive to New Portland.

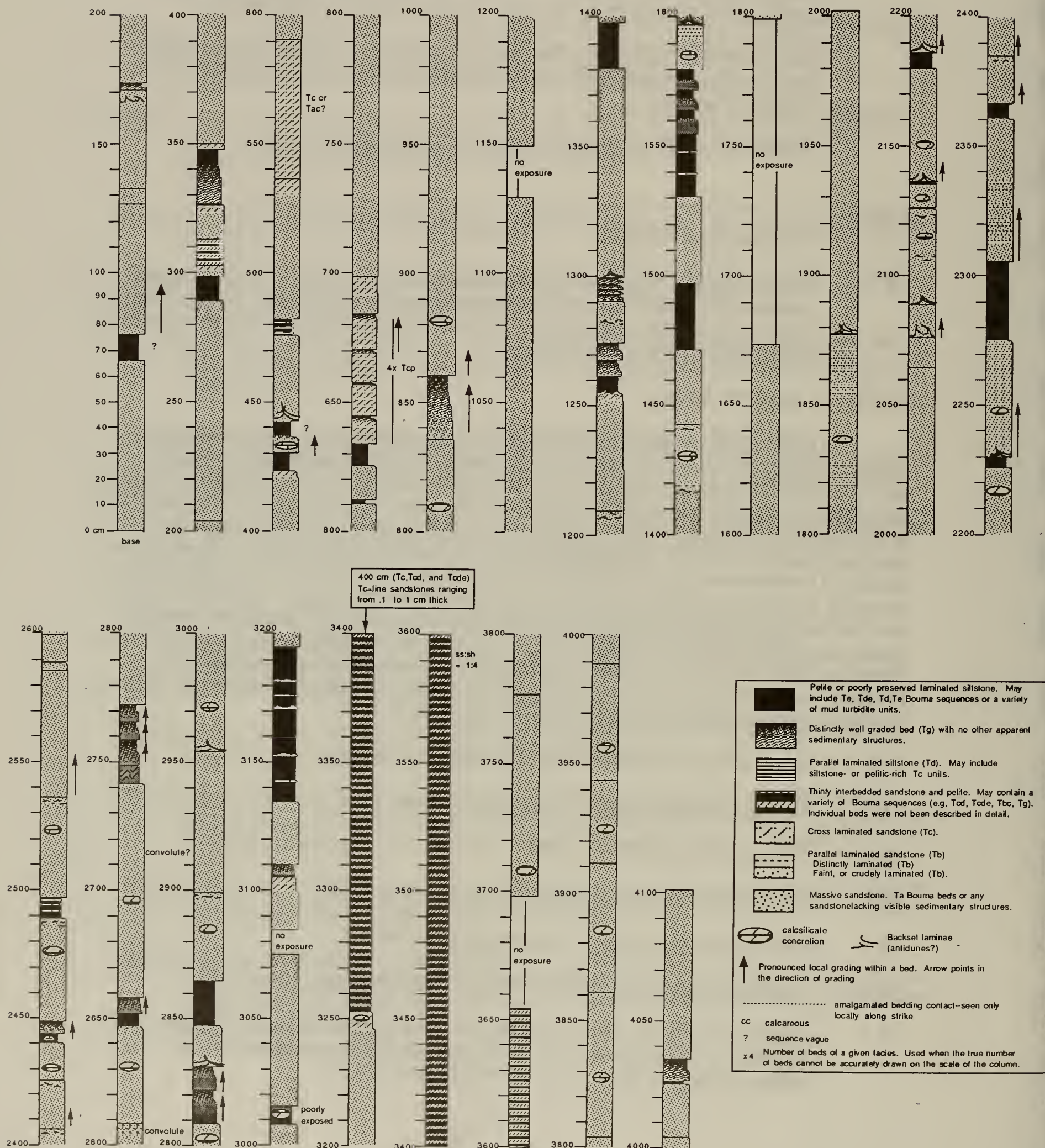
50.4 Turn east onto Route 146; drive to East New Portland.

54.7 East New Portland. Park along the road where Route 146 turns sharply left (N) to cross the Carrabassett River. The exposures of interest are along the southeastern bank of the river. Watch for poison ivy.



## Bradley and Hanson: Madrid Formation

Figure 4. Partial measured section of the Madrid Formation at its type locality along Sandy River. This section of Facies B amalgamated sandstones directly overlies the thinly bedded, more calcareous lower Madrid Formation (exposed in Saddleback Stream). The key applies to all measured sections.





## Bradley and Hanson: Madrid Formation

### STOP 2 - East New Portland

This is a good example of Cowan's (1985) Type I melange, consisting of variably disrupted turbidite beds and bed fragments of Madrid aspect, surrounded by a foliated metapelite matrix. Web structure is pronounced in sandstone blocks. East New Portland lies within the contact aureole of the nearby Lexington pluton (Figure 1). The melange is intruded by pegmatite dikes associated with the batholith. This pluton also cuts regional Acadian folds, and its contact aureole overprints regional biotite-grade metamorphism. An Rb-Sr isotopic age of 400 Ma for the Lexington (Gaudette and Boone 1985) brackets the age of older events. Deposition of the Madrid Formation, disruption of Madrid protoliths into melange, and Acadian regional deformation and metamorphism all must have occurred during the Pridolian-Gedinnian interval (about 414 to 401 Ma).

- 54.7 Drive north on Route 146, crossing the Carrabassett River.
- 54.85. Turn right (NE) onto River Road (shortcut).
- 56.75 Turn right (S) onto Route 16. Proceed slowly for about 0.2 miles and pull over at large roadcut.

### STOP 3 (Optional) - Contact between Madrid and Smalls Falls Formations

In roadcuts on the left (NE) side of the highway, sulfidic, rusty-weathering metasandstones and metapelites of the Ludlovian Smalls Falls Formation are in contact with purplish-weathering metasandstones of the Pridolian-Gedinnian Madrid Formation (eastern facies, originally designated as Fall Brook Formation by Pankiwskyj 1979). According to Pankiwskyj (1979, p. 19) the contact is a 2 meter thick melange zone; he interpreted it to be a premetamorphic, soft-sediment slump surface.

- 57.0 Continue on Route 16 to North Anson.
- 64.25 North Anson village. Turn right (S) on Route 201A. Cross the Carrabassett River.
- 64.4 Turn around and park on the northeast side of the road (Figure 5) where there is a very wide, paved shoulder.

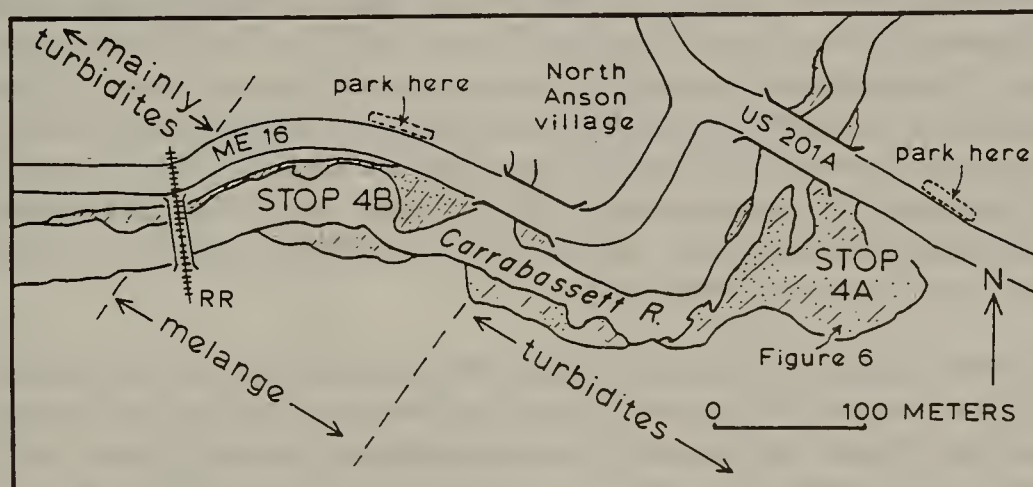


Figure 5. Sketch map of outcrops (stippled with strike lines of bedding and/or melange foliation) along the Carrabassett River in North Anson.



## Bradley and Hanson: Madrid Formation

### STOP 4A - North Anson Bridge

Cross to the west side of Route 201A and descend the embankment. About 100 meters south of the river is a series of recent exposures in which characteristic sedimentary features of the Madrid Formation are locally well displayed. A partial measured section (Figure 6) shows the abundance of amalgamated sandstones (Facies B). The bases of thick sandstone beds show characteristically rapid grading; backset laminae are also present. Younging direction, inferred from grading and backsets, is to the southeast. The outer-fan depositional setting was probably comparable to that at Stop 1.

64.4      Retrace route to center of North Anson.

64.55     Turn left on Route 16.

64.75     Park in front of the Public Library just beyond a bridge over Mill Stream, a tributary of the Carrabassett River. Climb over the guardrail and descend a concrete retaining wall to the river. For less agile participants, there is an easier but longer way down about 0.1 miles upstream, beneath the railroad bridge; the latter route involves some disagreeable bushwhacking in order to reach the best outcrops just below the library.

This will be our lunch stop. Food is available at the general store in North Anson.

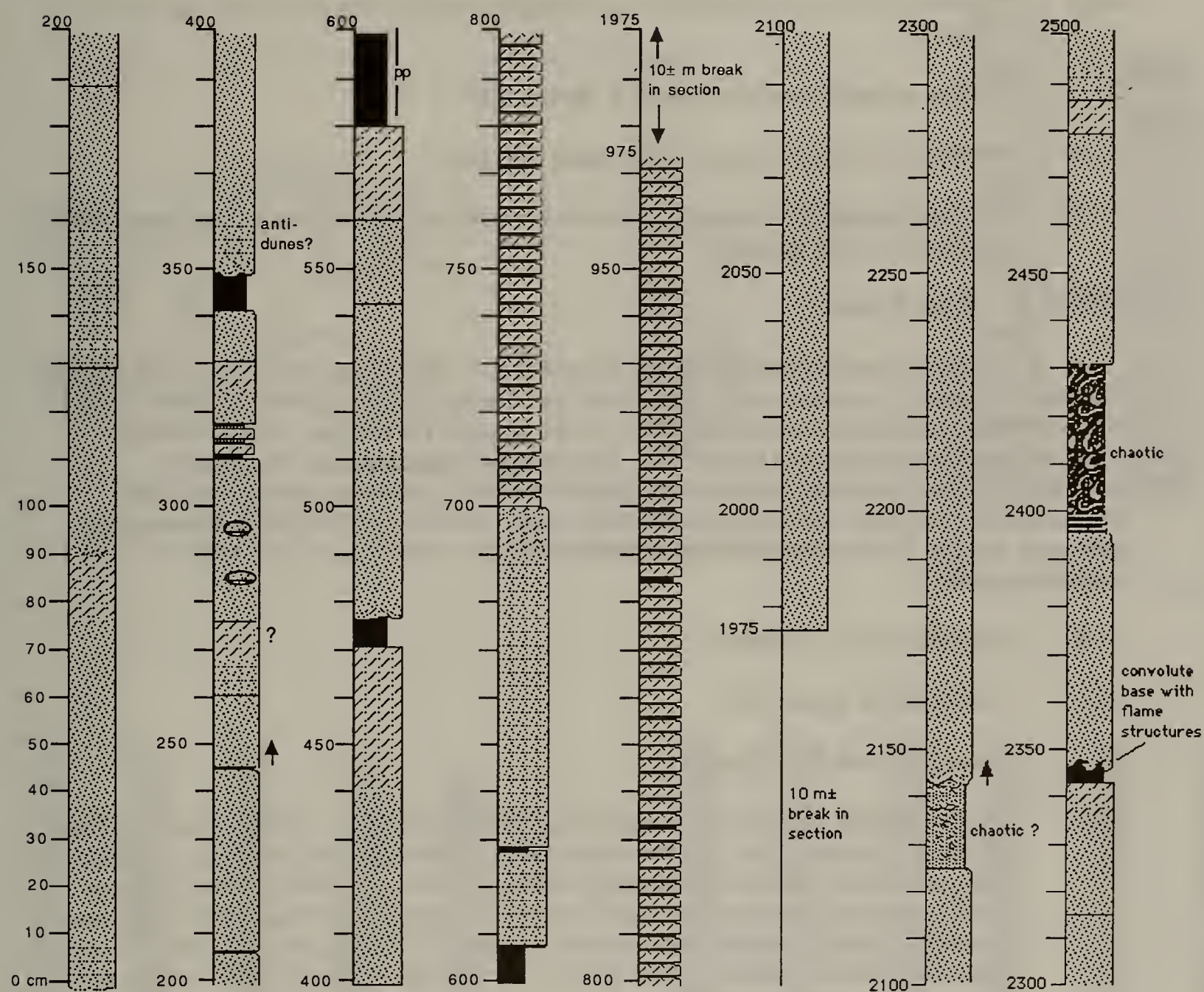
### STOP 4B - North Anson Melange

Exposures along the Carrabassett River from Mill Stream to the railroad bridge (Figure 5) are mostly melange. The melange is a stratabound package about 30 meters thick, bounded below and above by somewhat disrupted, but more nearly intact turbidites. The lower and upper boundaries of the melange zone are gradational; there is no evidence for a welded (sedimentary) upper contact such as would be expected if the melange were a slump deposit. The melange consists of disrupted layers and fragments of metasandstone, enclosed in a matrix of metapelite. Fragmentation of sandstone layers had progressed to varying degrees when deformation ceased, leaving a spectrum ranging from essentially intact bedding, to web structure, to partially dismembered but still recognizable layers, to isolated fragments with no obvious relation to neighboring fragments (Figure 3). Bedding is generally parallel to foliation in zones of relatively less intense deformation, but is commonly perpendicular to foliation in fragments in the more disrupted zones. The most prominent layering is a fragment foliation (Cowan 1985) defined by flattened metasandstone fragments surrounded by anastomosing films or zones of darker, finer-grained material; we interpret the foliation as a plane of bedding-parallel simple shear. Clasts are elongated subvertically in the flattening (foliation) plane. Byrne (1984) noted that bedding disrupted by web structure tends to break into prisms that initially are elongated perpendicular rather than parallel to the extension direction; consequently, the shear direction and sense are ambiguous.

The North Anson melange developed within partially lithified Madrid Formation turbidites. At the time of deformation, sand beds were sufficiently cohesive for fragmentation into blocks, as opposed to disaggregation into individual sand grains; at the same time, pelite was sufficiently mobile to flow into some available openings. Similar Type 1 melanges have been described by Byrne (1984) from the Ghost Rocks Formation, Alaska, where they have been attributed to early submarine thrusting in a subduction

## Bradley and Hanson: Madrid Formation

Figure 6. Partial measured section of the Madrid Formation at North Anson, south of the bridge over the Carrabassett River on US 201A. Turbidite facies here include Facies B (predominant), C, D, and F.





## Bradley and Hanson: Madrid Formation

complex. The melange at North Anson may represent what was originally a flat-on-flat thrust zone. Soon after the melange formed, the entire duplicated section was subject to regional Acadian folding.

- 64.75     Retrace path to intersection of Routes 16 and 201A in North Anson village.
- 65.0     Turn left (N) onto Route 201A, eventually crossing the Kennebec River, to Solon.
- 71.5     Solon village; turn left onto U.S. Route 201.
- 71.6     Turn left on Cross Street, drive about 200 feet
- 71.65    Park opposite first driveway (please do not block local traffic). Scramble down a short trail to the brook.

### STOP 5 - Fall Brook

Pankiwskyj and others (1976) designated this as the type section of the Fall Brook Formation (Figure 7), a name later abandoned by Moench and Pankiwskyj (1988) in favor of the informal designation "eastern facies" of the Madrid Formation. This is a classic thickening-upward sequence (Figure 7b), suggesting progradation in an outer fan environment. Southwesterly paleocurrent directions from cross laminae suggest that progradation was along strike toward the southwest. The field trip will visit the area of measured section 7e, characterized by abundant backset laminae and high ratio of sandstone to shale.

- 71.65    Retrace path to Route 201.
- 71.7     Turn left on Route 201.
- 72.3     Turn left onto Falls Road.
- 72.6     Park at a large cleared area overlooking a dam and what is left of Caratunk Falls on the Kennebec River. Archeological digs at the falls (for millenia a seasonal salmon fishing camp) have revealed artifacts as old as Middle Archaic (about 7500 years before present), as well as the much younger Woodland and Contact periods. A plaque commemorates the portage of Benedict Arnold's fleet of leaky bateaux here in October, 1775, during his campaign against Quebec in the early days of the Revolutionary War.

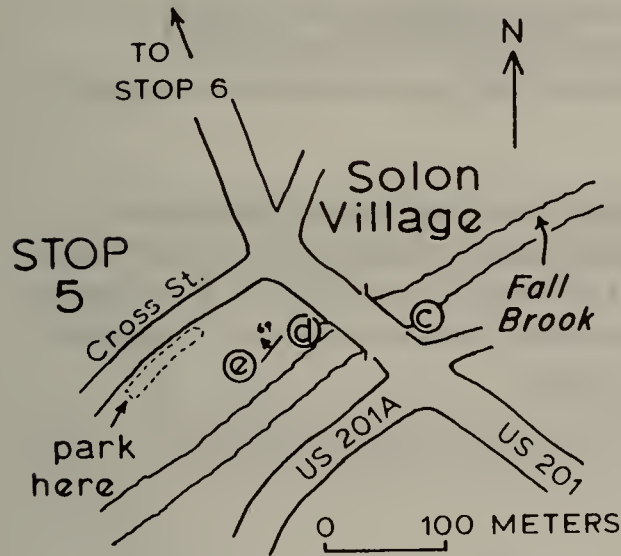
Walk downstream about 500 feet along a dirt road. Where it ends, follow a rough path to the right (watch for poison ivy) about 100 feet to large exposures along the river. Initials carved in the outcrop near its downstream end read "BA 1775".

### STOP 6 - Arnolds Landing

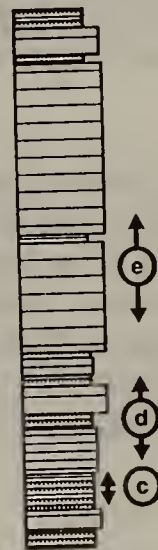
This stop features excellent exposures of the uppermost Madrid and lowermost Carrabassett Formations, and a passable exposure of the conformable contact. Figure 8 shows a detailed measured section across the contact and the transition from submarine fan to slope facies; the section begins at the river. The upper Madrid Formation consists of partial to complete Bouma sequences (Facies C and D) and amalgamated sandstones with

# Bradley and Hanson: Madrid Formation

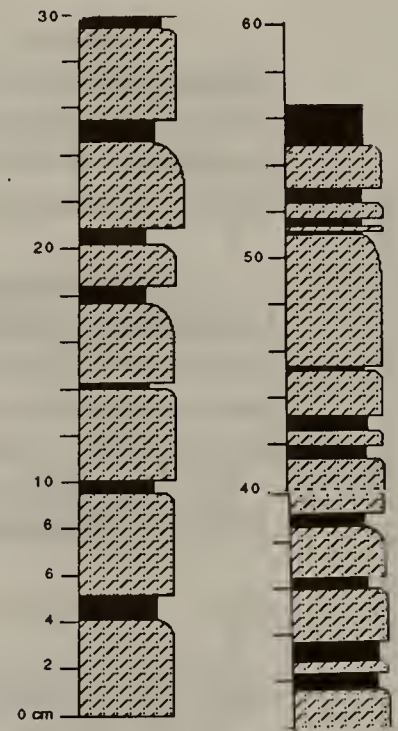
Figure 7. The outcrop belt at Fall Brook in Solon was the designated type section of the now-abandoned Fall Brook Formation (Pankiwskyj and others 1976); it now serves as a reference section for the informal "eastern facies" of the Madrid Formation. (a) Location sketch map. (b). Summary section showing thickening upward sequence. Letters show stratigraphic position of detailed measured sections (c, d, and e).



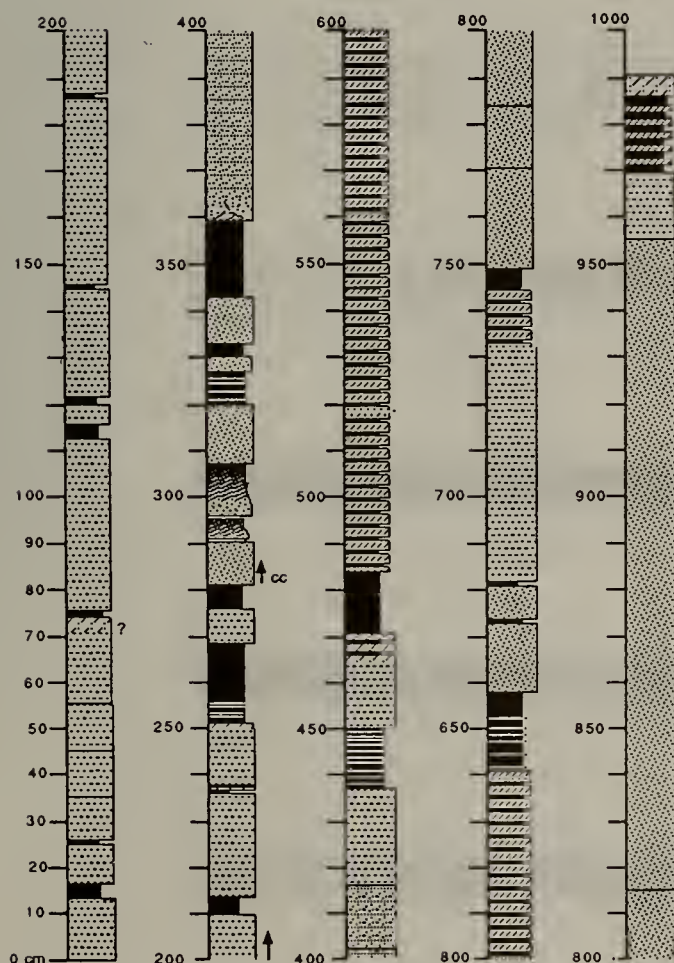
(a) Location map



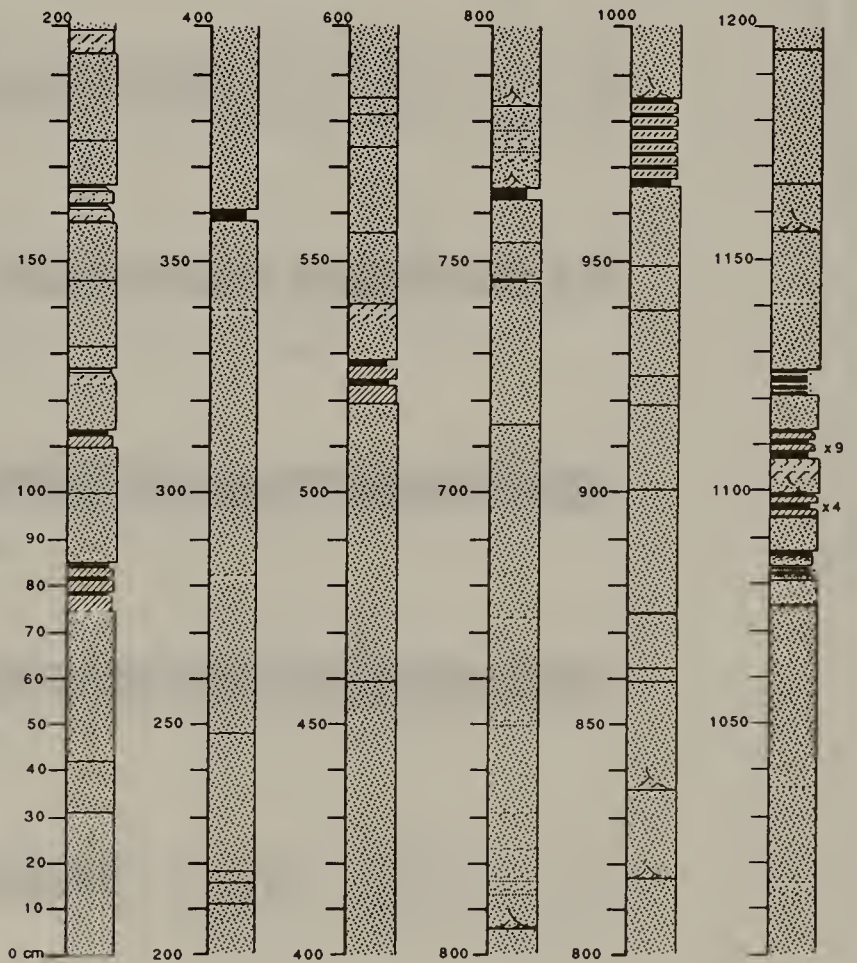
(b) Summary section



(c) Detail at location c



(d) Detail at location d



(e) Detail at location e



## Bradley and Hanson: Madrid Formation

erosive bases (Facies B). Channeling suggests a more proximal (e.g. mid-fan) depositional setting than the outer fan facies at Stops 1, 4, and 5. Paleocurrents obtained by single-tilt correction of cross laminae in Facies D turbidites suggest sediment transport toward the southwest, subparallel to strike. Together, this suggests fan progradation toward the southwest.

A slump horizon can be seen about 10 meters from the water's edge; contractional deformation in the slump mass suggests that it may have accumulated near the base of a slope. Near the water's edge, an early fault with ramp-on-ramp and flat-on-ramp geometries is interpreted as a contractional structure that probably predated or accompanied regional deformation.

The abrupt depositional contact between the Madrid and overlying Carrabassett Formation is poorly exposed in the woods along the access trail. Good exposures of the lowermost Carrabassett Formation are found a few tens of meters downstream. Bedding in these metapelites is marked by thin siltstone laminae, some of which are graded. They are interpreted as mud turbidites deposited on a submarine slope.

- 72.6      Retrace path to North Anson.
- ~80      Follow Route 201 to Anson
- ~85      Turn right on Route 43. Drive to Farmington where trip mileage began.
- ~105     **END OF TRIP**

# Bradley and Hanson: Madrid Formation

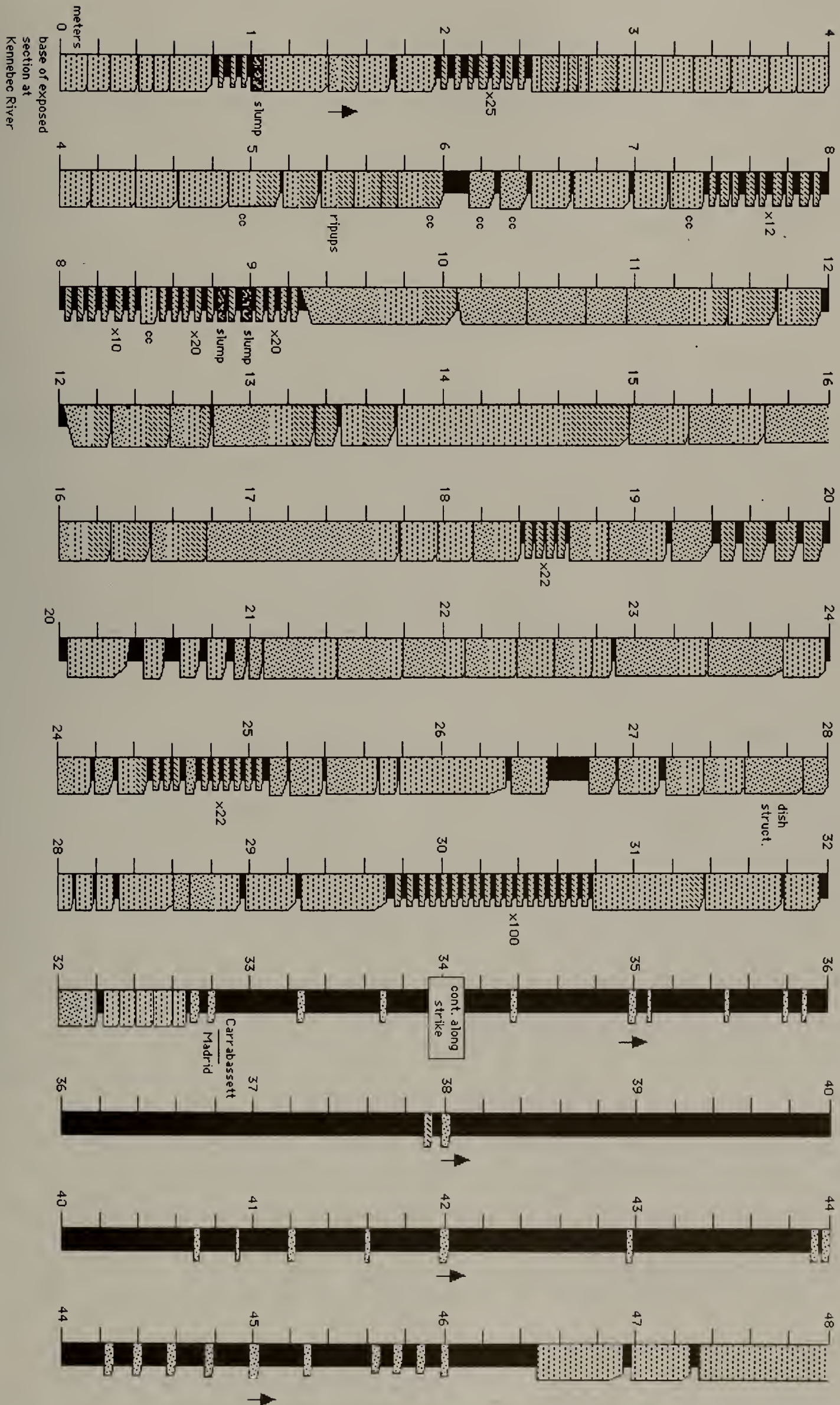


Figure 8. Partial measured section at Arnolds Landing, showing the uppermost eastern facies of the Madrid Formation, the lowermost Carrabassett Formation, and the abrupt depositional contact between.





### ENVIRONMENTAL GEOLOGY ALONG THE SANDY RIVER, FARMINGTON, MAINE

Field Trip Leaders: Dr. Tom Eastler, UMF  
Mr. Andy Buckland, UMF  
Mr. Joel Sproul, UMF

#### INTRODUCTION

The Sandy River is like a sleeping giant. When it awakes on occasion and cries out, repercussions are felt throughout the system. But even as the giant sleeps, man and his machines are altering the environment about this mighty river, and in so doing are ensuring that in each subsequent awakening, ever more destructive activity and potential alteration of the very course of the river will ensue.

Today's field trip will cause us to visit but a very few of the areas of the Sandy where dynamic activity has recently, is now, or soon will be occurring. The following material will set the stage for your excursion to the sleeping giant.

#### THE LIFE-BLOOD OF THE GIANT (the watershed)

The Sandy River watershed drains about one-third of Franklin County's one million plus acres. At Farmington Falls, near the southern edge of our field trip area, the Sandy drains some 420 square miles of watershed; at Mercer near its confluence with the Kennebec it drains 514 square miles. Nearly all of the Sandy River lies within Franklin County. The headwaters of the Sandy are in the northwest corner of State, in the Township of Reddington and Town of Madrid, and it flows in a southeasterly direction through the towns of Phillips, Avon, Strong, Farmington, New Sharon, and Mercer; it traverses some 69 river miles until it empties into the Kennebec River in the Township of Starks (York, 1980).

Although the average annual discharge at its mouth is about 980 cubic feet per second (cfs), the extremes to which the Sandy is subject are notable. At Mercer in the driest of years the flow has been as low as 32 cfs (Sept. 22 thru 26, 1939), and as high as 51,100 cfs (1 April, 1987) (Bartlett, et. al., 1989).

Overall the Sandy River watershed is of rural character, although small urban-like centers are located at Madrid, Phillips, Strong,



Wilton, and New Sharon, with Farmington being the largest and most urban area. The Watershed is more than 85% forested, with the remainder of the area occupied by agricultural lands and urbanized areas. Over 90% of agricultural productivity in Franklin County falls within the wide fertile flood-plain intervale land of the lower reaches of the watershed (York, 1980).

The watershed borders on both the Northern and Southern Interior Climatic Zones and has average daily temperatures ranging from about 18 degrees F in January to 69 degrees F in July, although extremes of less than -40 degrees F and greater than 100 degrees F are not uncommon. The average annual precipitation is about 45 inches, which includes the water equivalent of nearly 100 inches of snow (USDA/SCS, 1976).

#### THE AWAKENED GIANT'S SCRAPBOOK (Flood History)

The Sandy River has had a very long and interesting history of flooding. Stream gage readings date back to October, 1928, but historical records and oral tradition take us all the way back to 1785, some 9 years before the town of Farmington was incorporated (York, 1980). York reports 44 recorded floods and freshets from 1785 to 1960. Eastler (1989) records an additional 51 floods from 1961 to 1989, giving a grand total of 95 recorded flooding events in a 204 year history, or about one newsworthy flood in Franklin County every 2 years. Figure 1 shows just a few of the newspaper headlines that have appeared over the years.

In terms of severity, the 1987 flood (51,100 cfs) ranks first, the 1936 flood (38,600 cfs) second, and the 1953 flood (36,900 cfs) third. It is noteworthy that the most damaging floods on the Sandy have usually been associated with heavy rainfall that occurred during the winter or early spring months. As a result of numerous ice jams, most of these floods reached artificially high stage levels with respect to the actual runoff recorded. Thus, although flooding is usually a rain driven phenomenon, this has usually not been the case on the Sandy River. The April Fool's flood of 1987 was a happy exception to the above stated rule; and indeed what a blessing that was. If ice jams had been associated with that flood, the several million dollar's damage in Farmington alone could have reached many times that figure. Quite possibly every bridge from Fairbanks to Farmington Falls would have been washed away.

#### THE ENVIRONMENTAL EFFECTS OF FLOODING

The Sandy River has been a textbook example of a rural hydrograph up until recent years (the 1970's). With the exception of severe



March 17, 1936

## THE SANDY GOES ON A RAMPAGE ALONG WITH OTHER MAINE RIVERS

Heavy Rainfall on Thursday Caused Great Loss in Kennebec County—Franklin County's Loss Comparatively Small

The Sandy River was one of the worst off in the county. The Farmington School was flooded, water standing 1 foot deep on the first floor. It rained in such a way that the water was unable to remove any of the mud that had been on the roads. The filling station near by was

June 16, 1942

## Sandy River Goes on Somewhat of a Rampage

Cloudburst Above Phillips and Torrential Rain All Along the Line Render Some Roads Impassable and Cause Flooding of Corn Fields

Sandy River was swelling and continued to rise for twelve hours after cloudburst. Heavy rains. This condition will cause a great amount of damage.

March 18, 1947

## Farmington Citizens Fear Flood as Sandy Overflows

Highway Goes Under Foot of Water Near Devil's Elbow---Ice Jam Breaks from Heavy, Warm Rain or Big Thaw May Cause Great Damage

Farmington residents were this past week-end watching the rising waters as the result of a heavy, warm rain or a big thaw, extensive damage will be done. To add to the danger, many

## Sandy River Goes on Rampage-- Damage Estimated in Thousands

Series of Electrical Storms Brings Torrential Rainfall of 40 in. in 24 Hours --- Heavy Loss Intervale Crops---Many Cellars Are Flooded

June 18, 1943

Some Highways Were Impassable --- Will Cost Farmington \$2000 to Repair Road Damage --- Bolt Strikes Transformer

A series of electrical storms, accompanied by heavy rain, starting at Walton's Mill, has caused much damage to many belongings of residents.

FEBRUARY 7, 1947

## Unseasonable Rainfall Sends Rivers, Streams on Rampage

Ice Jams, Wednesday, Sent Water Coursing Over Highways and Lowlands, Blocked Traffic, Marooned People in Homes and Business Places and Flooded Cellars

A few warm days and several hours of rain have started the rivers receding back to normal. Ice was jammed for over a mile from

FIGURE 1. A glimpse at a few headlines seen in local papers over the years concerning flooding on the Sandy River.



floods in 1785, 1820, 1855, 1869, 1928, and 1936, the river has risen slowly in response to rainfall (but not ice jams) and decayed to baseline equally as slowly. Floods in 1947, 1950, 1951, 1954, and 1959, seemed to suggest a more frequent recurrence of newsworthy floods than in the past. Floods in 1960, 1963, 1964, 1966, 1968, 1969, 1973, 1975, 1978, 1979, 1983, 1985, 1987 definitely demonstrated that the system was somehow changing, both in the severity of more common floods, and the repetitive frequency of those floods.

It seemed that flood stage was being reached sooner and severity was greater, while flood duration was less, except in the case of ice jams. Could this be the transition into the Urban hydrographic situation??? It seemed so.

An evaluation of the damage levied on the the once nearly stable banks of the river, suggested that bank erosion was on the increase, and that large long-standing bank-lining shade trees were on their way out by increased erosion. The usual culprit for such drastic changes in the equilibrium of the river is development. Increased clearcutting deep in the watershed coupled with increased paving and building was bound to exacerbate the situation and hasten the change to an urban hydrograph.

### The Role of Ice in Flooding

One contributory factor resulting in artificially high discharge amounts as well as increased flood frequency has been the build up of gravel bars in the river. Gravel had been removed at will since before 1785, and on an as needed basis for the health of the river bars would be removed to keep the historical channels open. But in 1975 it was determined by the State that river gravel could not be mined from the river channel anymore, and permits to remove sand and gravel from the periphery of the river were very hard to come by. The Department of Inland Fisheries and Wildlife were the adjudicators of the permit system, and the onus was on the permit seeker to demonstrate that adverse environmental effects would not occur if gravel removal were allowed. Stymied by the numerous and cumbersome bureaucratic stumbling blocks, virtually all gravel removal operators moved inland to the floodplain areas and began removing fertile topsoil to expose the ancient river run gravel underneath; this gravel was under no jurisdiction and could be removed at will. Much potentially valuable intervale farmland was lost to this process and replaced with a stark gravel pit environment which after abandonment became a relatively unproductive intervale eyesore.

As a result of the lack of removal of river channel deposits, bars began to build up unimpeded by the mining practices of man for nearly four years. Subsequent floods, in particular the 1978 flood where a great loss of intervale farmland was linked to ice build up on the newly grown gravel bars in the Farmington area,

finally caused the State to reassess its policy and allow renewed river gravel removal by application of a more liberal permitting policy, although dredging from below the waterline was still not normally allowed. It is apparent now that even the more liberal permitting has not substantially reduced the amount of gravel clogging the Sandy River because much of the gravel still resides below the normal low water mark, and still catches ice during the Spring runoff.

Unfortunately much damage was done by the years of non-removal, and bar build-up in the Sandy river even now is at a near critical stage. Compounded by increased surface runoff, the banks of the Sandy are receding in places up to 10 feet per year (30 feet per year if you believe one unsubstantiated claim). Agricultural land is being eroded away at frightening rates, and a less than optimal permitting procedure is hindering the expeditious removal of the most threatening bars.

Meanwhile floating ice and ice jams combined with unusually high discharge rates (which don't seem to be so unusual anymore) are continuing to take their toll on our precious land.

#### Pre-emptive Mitigation

All of the above taken together suggest that some activity should take place that might lessen the toll that is being taken on fertile intervale floodplain land. Little of immediacy can be done to stem the rapidly rising waters from rain driven runoff. But much can be done to minimize the damage done by ice build-up, ice-jams, and ice-dams. Since river encroaching gravel bars are clearly clogging the rivers and redirecting them as well as catching ice, removal of specific gravel bars would be highly recommended in this situation.

In 1987 the Franklin County Soil and Water Conservation District took an active part in trying to identify troubling gravel bars to be earmarked for possible removal. Under a challenge grant from the Maine Soil and Water Conservation Commission to the Franklin County Soil and Water Conservation District, funds were made available for study of the nature of the gravel buildup in the Sandy River. Those studies are ongoing now. Preliminary investigation has identified some 147 gravel bars within the confines of Franklin County, many of which should be removed or reduced in size forthwith.

The stops in this field trip are designed to illustrate some of the problems associated with the build up and continual growth of sand and gravel bars within the Sandy River. Since final conclusions have not yet been drawn regarding the most appropriate mitigation procedures, the field trip participants will be asked to input their ideas during the trip in order to ensure that many points of view will have been discussed prior to the establishment of final recommendations.



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## ITINERARY

ASSEMBLY POINT: Parking lot of the University of Maine Farmington Geology Center, 120 Main Street, Farmington, Maine.

MAP: see Figure 2

Proceed southerly on main street to the intersection of Rte 2.

### Cumulative mileage

- 0.3 Left onto Rte 2 east
- 9.5 Right after Green Bridge onto Rte 134 south, New Sharon
- 9.9 Left onto Smith Road
- 11.1 STOP 1 Smith Quarry

Granite quarry from which 50,000 yds of granite were removed to be used as rip rap for the New Sharon Gorge project. This will be a short stop to examine the nature of the area bedrock and to appreciate the scenery from the Smith Farm.

- 12.3 Right onto Rte 134 north
- 12.7 Left onto Rte 2 west (while on bridge be sure to look downstream (right) to the left side of the Sandy River to view the protected embankment).
- 12.9 Right onto Rte 134 north
- 13.1 STOP 2 Sandy River Gorge
- This will be a brief stop along the top of the protected embankment. This stop is about 100 feet above the river.
- 13.5 STOP 3 Rip Rap, Sandy River Gorge Embankment
- After a short hike along the old haul road, we will examine the finished product of about one year's work by the Maine Department of Transportation at a cost of about one million dollars. Discussion of environmental impacts will follow.
- 16.2 Right onto dirt road to Sandy River Estates.
- 16.9 STOP 4 Sandy River Estates
- Here we will examine erosional processes ongoing which threaten to unequally disrupt the land ownership in the Sandy River Estates. Some owners will be relatively unaffected, others stand to lose much land. The field participants will be queried as to how they feel this situation should be handled.
- 17.6 Left on Rte 134 to Rte 2
- 20.8 Right on Rte 2 west
- 23.7 Left onto dirt road into cornfield accross from New Sharon Motel.
- 24.0 Right through fence and follow field road to right.
- 24.4 STOP 5 Davis Island
- Within the last year a 40 acre bulbous peninsula has been converted by the forces of nature into a 40 acre island. This meander cutoff phenomena is not often viewed in the making, and participants will be treated with the knowledge that they are witnessing a significant geologic event.
- 24.8 Left through gate and back to Rte 2
- 25.1 Left onto Rte 2 west.



26.2      STOP 6    Hiscock House

During the April Fool's flood of 1987 Thelma Hiscock's cellar collapsed into the Sandy River, nearly taking the rest of the house with it. A federally funded project (approx \$20,000) restored the back yard and the firm footing onto which their house could set. Subsequently the house was restored and sold to a third party.

26.3      Left onto Farmington Falls road to Brad Moore's house.

26.4      STOP 7    Brad Moore's house

Several houses here were severly flooded. The major threat was from water flooding them from behind rather than from in front. The State Rte 2 roadbed acted like an earthen dike (dam) and diverted the flow of flood waters from upstream accross several properties to an outlet downstream; the outlet just happened to be at this stop.

26.5      Left onto Rte 41 thence across bridge

26.6      Right just after bridge onto Rte 156

27.0      Stay on Rte 156 and cross green bridge (right)

27.3      STOP 8    Schultz's Field

This 180 degree change of river course has been altering the local scenery to some degree since 1780. Rates of erosion here have accelerated over the last 20 years or so, and have reached 10 feet per year just upstream as of several years ago; those rates have diminished some over the last several years, but seem to have accelerated recently downstream at this site.

Several very interesting environmental activities are currently underway here. The gravel bar accross the way (Pillsbury's bar) has not yet been removed this year due to permitting red tape thus ending a 15 year or longer annual removal cycle. Meanwhile the State DOT has just completed rip-rapping part of the road embankment while at the same time removing topsoil from the very embankment that is eroding away into the river.

27.4      Right onto back Falls road.

27.5      STOP 9    Hines' house

At this stop in the late 1970's a house was situated at just about the current location of the steep

embankment. Rates of erosion of their back yard (now non-existent) reached ten feet per year for about 4 years prior to the removal of the house. Subsequently the rates have been reduced to about a foot per year. An interesting story of pre-emptive flood mitigation will be discussed at this stop.

31.5 Right onto Rtes 2 east and 4 north.

32.5 Left after crossing center bridge across the Sandy R.

32.8 Sharp Right onto Lower Main Street to Geology Bldg.

32.9 LUNCH STOP

There will be a thirty minute lunch stop at the geology building. Several fast food joints are within a one minute walk from the building. Inside the building will be a poster board display of some of the features already discussed today as well as of those to be visited this afternoon.

0.0 Left onto Lower Main Street.

0.3 Right onto Rte 2 west

1.3 Left into Pike's vegetable stand

1.6 STOP 10 Pike's Field

This stop is a prime example of how not to address environmental problems. David Pike was (and still is) a very practical and efficient farmer who tweaked every possible ounce of productivity from the Hadley silt-loam soil on the 15 acres of intervale farmland here. The year of the great disastrous, ice-generated flood at this place, Pike had just been awarded the "Conservationist of the Year" award from the local Soil and water conservation district for his conservation practices.

The story of personal loss and societal tragedy to be told from this stop will not soon be forgotten. To be seen will be the loss of 3 acres outright to a new "meander cutoff", the isolation and subsequent demise of 12 acres of highly productive soil, and the vestige of a losing effort to save the island for agriculture.

This stop holds many lessons for the scientist and environmental manager alike.

1.9 Right onto Rtes 2 east and 4 North

2.3 Left onto Town Farm road.



- 5.8 Left onto Rte 4 north
- 10.7 Right into field to Crandall Bar
- 11.0 STOP 11 Crandall Bar
- More than a year passed by in obtaining a permit to remove the gravel from this bar. The permit was just recently issued (in a modified form from the desired removal strategy) too late to remove gravel this year, so another winter shall pass before extraction begins. Participants will discuss the choice of removal boundaries at this site.
- 11.3 Left onto Rte 4 south
- 14.7 Left into Peter Tyler's gravel bar
- 15.0 STOP 12 Tyler gravel bar
- The opposite bank erosion at this stop is classic. Over a hundred feet high and several hundred feet long, this bank dumps in about 4,000 yds of sand, silt, and clay for every yard of embankment removed at the toe of the bank; all of this material makes its way downstream. Also, about 6000 cubic yards of gravel were added to this gravel bar this summer alone. Removal strategy at this bar will also be discussed by the participants.
- 15.3 Left onto Rte 4 south
- 17.1 STOP 13 Fairbanks Bridge
- A quick look at the replacement bridge for the old Fairbanks bridge that washed away during the April 1987 flood. Note the embankment protection features.
- 18.7 Right into Lambert's Trailer Park
- 18.9 STOP 14 Lambert's Trailer Park
- Only one trailer remains of the eight that were here before the 1987 flood. All trailers were totally flooded and one was whisked away by the flood waters. The participants will discuss how to deal with placement of such structures.
- 19.1 Right onto access road to Rte 4 south
- 19.2 Right onto Rte 4 south
- 19.8 Right into Pike Glacier Ledge Farm (turn at LUHTA sign)

20.0 Right toward river

20.3 STOP 15 Pike's gravel bar

Yet another River/Farm interaction. David Pike continues to battle the elements even from this vantage point. A discussion of strategy for river management will conclude at this stop.

20.6 Left up hill

20.9 Right onto Rte 4 south

22.0 Left into Learning Center parking lot at UMF

22.1 Left onto Lower Main Street

22.2 FINAL STOP AT GEOLOGY CENTER





FIGURE 2. Location map for Sandy River field trip.



# BEDROCK GEOLOGY OF THE SOUTHEASTERN MARGIN OF THE CONNECTICUT VALLEY - GASPÉ SYNCLINORIUM IN SOMERSET COUNTY, MAINE

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## INTRODUCTION

The area encompassing the Penobscot Lake, Sandy Bay, and Seboomook Lake 15 minute quadrangles, collectively referred to as the Penobscot Lake region, lies on the southeastern margin of the Connecticut Valley - Gaspé synclinorium, near its junction with two other major tectono-stratigraphic sequences -- the Boundary Mountains anticlinorium and the Moose River synclinorium -- from which it is separated by major faults (Figs. 1 and 2). Sedimentary rock units of the Penobscot Lake region are interpreted as turbidites thought to have been deposited on the slopes of a basin, possibly extensional, following the Taconian orogeny and before the Acadian. They record a successive increase followed by a decrease in the distance to source region, and may reflect a change in provenance direction as well. The units have been moderately to tightly folded, thrust to the northwest, and metamorphosed to greenschist facies during the Acadian orogeny. South and southwest of the area, intensity of Acadian deformation and rank of metamorphism rapidly increase (Osberg et al., 1985). Beyond the northwestern margin of the Connecticut Valley - Gaspé synclinorium the effects of the Taconian orogeny intensify (St-Julien and Hubert, 1975).

## REGIONAL GEOLOGIC SETTING

### Connecticut Valley - Gaspé synclinorium (CVGS)

The CVGS extends from the eastern tip of the Gaspé Peninsula, southwestward across southern Quebec and northwestern Maine, then south through eastern Vermont, and central Massachusetts and Connecticut. It contains a thick and varied sequence of marine clastic deposits of primarily Silurian and Devonian age which, for the most part, rest unconformably on Cambro-Ordovician rocks. Williams (1978) grouped the rocks in the synclinorium in his successor basin lithofacies, suggesting that the rocks were deposited on the destroyed margins of a former ocean. However, the successor basin model is complicated by extensional and transcurrent activity which may have contributed to the development of the basin.

### Boundary Mountains anticlinorium (BMA)

The BMA extends from just west of Moosehead Lake southwestward to the Maine-New Hampshire border where it joins the Bronson Hill anticlinorium. Some workers extend the anticlinorium as far north as the pre-Silurian exposure at Caucomgomoc Lake (Osberg et al., 1985). Early studies of the bedrock geology of the BMA (Billings, 1956; Green, 1964) established much of the stratigraphy of the Precambrian through Middle Ordovician rocks exposed there. At its lowest stratigraphic level is the Chain Lakes massif, which



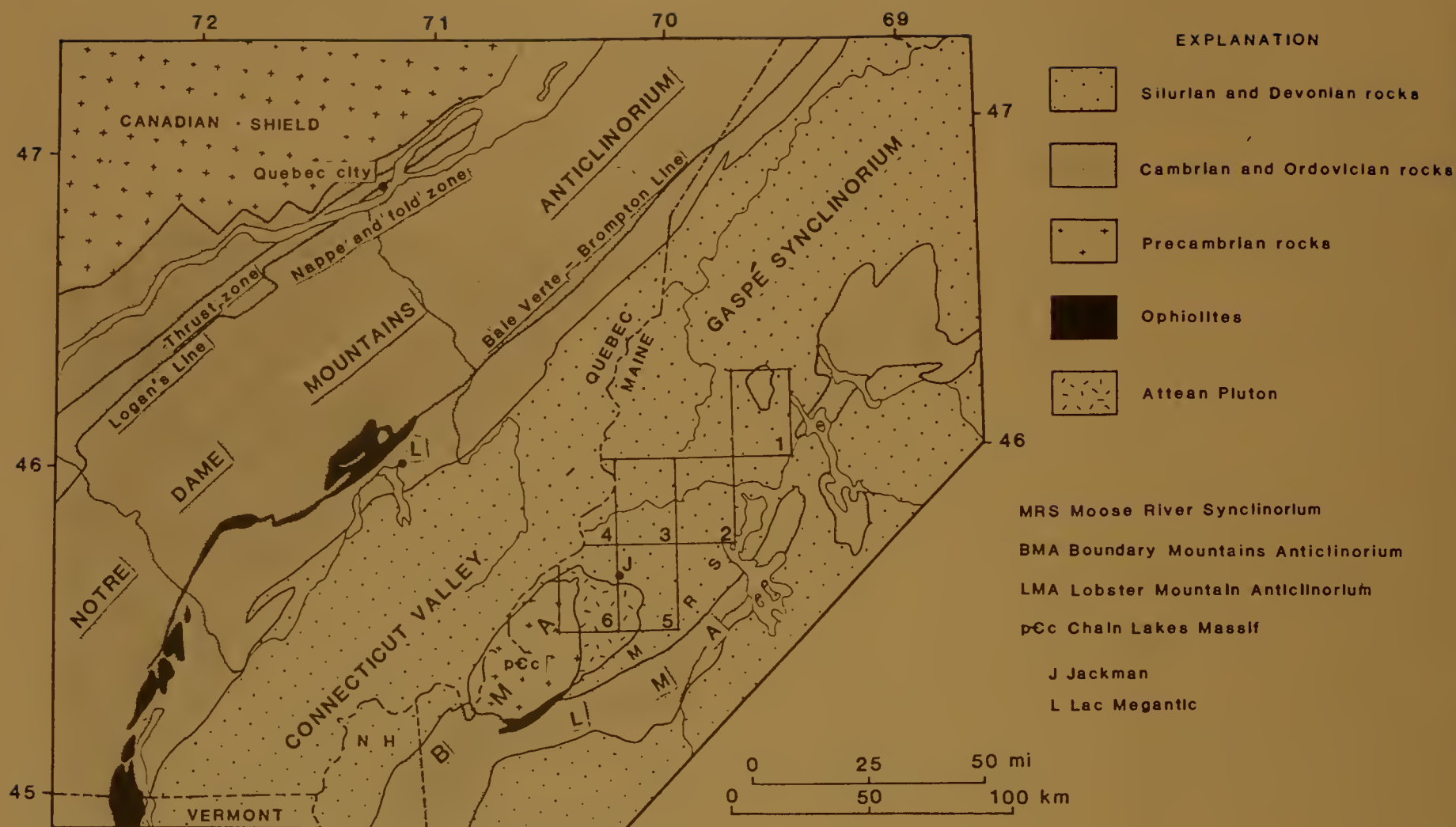


Fig. 1. Geologic outline map of northwestern Maine and southeastern Quebec. Number quadrangles are: (1) Caucomgomoc Lake, (2) Seboomook Lake, (3) Penobscot Lake, (4) Sandy Bay, (5) Long Pond, (6) Attean.

consists of a complexly deformed group of gneisses, granofels, and other high grade metamorphic rocks which are considered to be the basement of a micro-plate or terrane sutured to North America (Lyons et al., 1982; Osberg, 1978; Boone and Boudette, 1989). On the southeast of the massif is welded the Cambrian Boil Mountain ophiolite sequence (Boudette, 1982). Succeeding this is a Cambro-Ordovician accretionary mélange and flysch sequence (Boone, in press). On the eastern side of the BMA the pre-Silurian rocks are overlain unconformably by Silurian strata of the Moose River synclinorium (Albee and Boudette, 1972).

### Moose River synclinorium (MRS)

Study by Boucot and Heath (1969) resulted in definition of most of the Silurian and Devonian rocks in the MRS. The rocks predominantly are mildly metamorphosed, fossiliferous sandstone and mudstone, and are again representative of the successor basin lithofacies defined by Williams (1978). The Acadian orogeny tightly folded the strata into a series of minor doubly plunging anticlines and synclines. Recorded in the sedimentary rocks of the synclinorium are the early stages of this orogeny. Where the MRS meets the CVGS, the stratigraphic sequence is essentially conformable though modified by major faults.

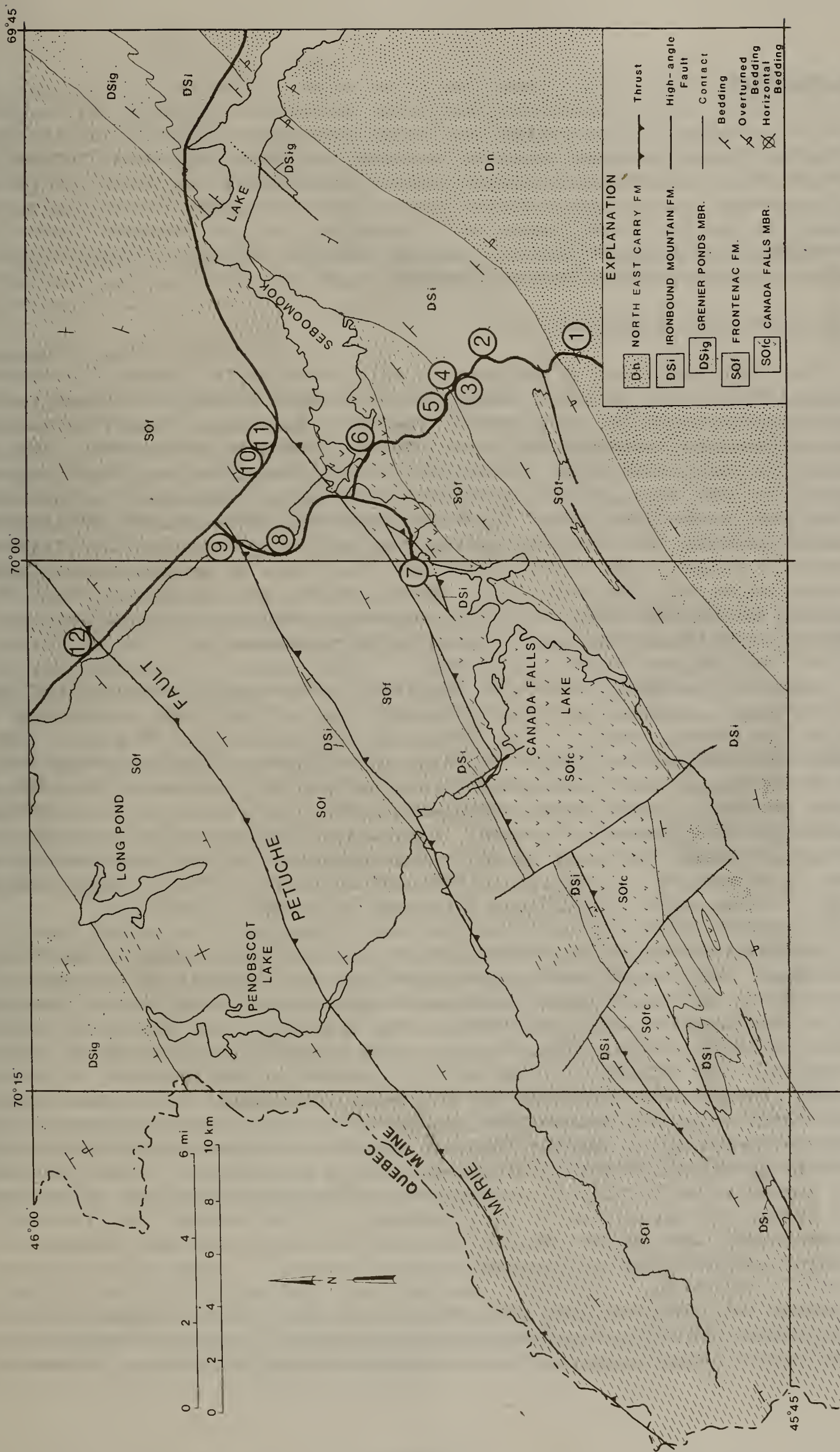


Fig. 2. Generalized geologic map of the Penobscot Lake region showing fieldtrip stops.



## STRATIGRAPHY

In the Penobscot Lake region the units from youngest to oldest are the North East Carry Formation and the Ironbound Mountain Formation, both of which are members of the Seboomook Group (see Pollock, 1987, for justification of the Seboomook Group), and Frontenac Formation. The Ironbound Mountain includes the Grenier Ponds Member and is underlain by the Frontenac Formation. In addition to the main body of the formation the Frontenac contains one member, the Canada Falls volcanic member.

### North East Carry Formation

The Seboomook Group is the unit to which most of the rocks of northern Somerset County and those of the northwestern half of Aroostook County, Maine are assigned (Osberg et al., 1985). The North East Carry Formation now includes the type locality rocks of the former Seboomook Formation. Originally called the "Seboomook Slate" by Perkins (1925), the group was described as a variably sandy, dark gray slate with well developed cleavage. In their work on the MRS, Boucot and Heath (1969) redefined the formation and designated the rocks exposed at Seboomook Dam as the type section even though it is there atypically sandy. This redefinition has been the a primary source of stratigraphic confusion. Pollock (1987) summarized the historical aspect of the term and pointed out that many field workers currently adhere to Perkins' original definition which well fits most of the formation.

Rocks of the North East Carry Formation are recognized in a narrow belt which strikes southwestward from Seboomook Dam and continues across most of the southeastern quadrant of the Seboomook Lake quadrangle (Fig. 2). The entire southeastern section of the area is underlain with these rocks but they are difficult to trace with certainty to the south. In areas of the MRS where Boucot and Heath (1969) mapped Seboomook rocks, the extent and distribution of North East Carry rocks is unknown. Thickness of the formation can only be estimated because the upper contact is not exposed in the field area and because of structural repetition in the outcrop belt. Based on known outcroppings of these rocks the minimum thickness is around 1000-2000 m.

Cyclically bedded fine quartz-arenite and gray slate comprise most of this lithology. Beds range in thickness from 2 cm to 2 meters; the average is around 10 to 15 cm. Bedding is arranged in packages of uniform thickness such that significant variation is usually seen only when one compares outcrops across regional strike. This is a function of both the thickness of each package and the thickness of the section exposed in each outcrop. Exceptions to this general rule occur when an unusually thick section is exposed such as that at the base of Seboomook Dam, where the thickest beds are found. By contrast, alternating packages of thickly and thinly bedded units can be traced directly across the width of Seboomook Lake, a distance of 1.6 km (1.0 mi). This demonstrates relatively wide-spread uniformity of controls on sedimentation during the deposition of each package. On the average the lithology is 50% sandstone, but in the thicker beds the sandstone fraction is dominant and in thin beds the opposite is true.

Almost all beds are very well graded from fine- to medium-grained sandstone. They begin at sharp bases and grade smoothly upward to mudstone. There are no abrupt upper contacts of the sandstone fractions. Concomitant

color change is from buff at the base to dark gray at the top, which produces the varve-like appearance described by Boucot and Heath (1969). Color in the mudstone slate can vary irregularly from blue-gray to slightly greenish gray. About 13% of the sampled outcrops display convolute laminations which were probably the result of small-scale slumping. Other sedimentary features of note include cross-bedding, sole markings, and bedding-parallel laminations.

Boucot and Heath (1969) defined the age of the former Seboomook Formation in this area as late Gedinian which probably is applicable to the North East Carry Formation.

### **Ironbound Mountain Formation**

The Ironbound Mountain Formation is exposed in two major belts within the present field area: one in the northwest and the other on the southeast side (Fig. 2). These belts extend to the northeast and southwest beyond the Penobscot Lake region. Two minor strike belts bounded on their southeast margins by thrust faults are wholly contained within the region. The southeastern major belt thickens southwestward and is partly mapped into areas previously identified as Seboomook Formation. Northeastward this belt extends as far as Caucomgomoc Lake where it ends by facies transition.

The northwestern belt is underlain entirely by the Grenier Ponds Member. The extent to which this member can be traced to the southwest in areas mapped as Compton Formation (Marleau, 1968) is unknown. Areas along the northeast extension have been mapped traditionally as Seboomook Formation (Osberg et al., 1985).

Thinly bedded, well-cleaved, medium- to dark gray mudstone and siltstone slate is most characteristic of the Ironbound Mountain Formation proper. Beds range from 1 to 30 cm in thickness but averages around 10 cm. Bedding is usually regular in thickness and orientation. Beds are frequently graded and their bases sharp, giving the appearance of cyclicity often ascribed to Seboomook rocks (cf. Boucot and Heath, 1969).

In many outcrops of mudstone slate bedding cannot be distinguished resulting in a massive appearance for the unit. Where bedding is recognizable in rocks of this lithology it is usually marked by brown silt laminae about a millimeter in thickness at the base of each bed. Such beds usually show no grading. Siltstone slate beds are well graded from silt to clay grain size over the entire thickness, which is generally 10 cm or less.

Rocks of the unit are usually gray but may locally be olive-gray. This is especially true of the rocks near Ironbound Pond in the southwestern part of the area and is attributed to a higher chlorite content there than elsewhere in the formation. In the same area the rocks display a phyllitic sheen which is related to increased phyllosilicate content.

Two local variations in the lithology of the formation constitute minor percentages of the unit. Scattered with no apparent vertical regularity in the rest of the unit proper are thick-bedded, light gray, graywacke. These become more prevalent toward the northeast. The graywacke beds are typically massive, fine-grained, well sorted in framework constituents, and either show no grading or are poorly graded. Beds of this lithology usually are not found in groups and are not traceable laterally.



At several localities randomly distributed in the Ironbound Mountain are minor beds of matrix-supported conglomerate. Bedding is generally poorly defined and may be up to 2 meters in thickness. The matrix is generally of well cleaved, dark gray mudstone slate. Clasts are usually rounded and consist most commonly of slate and feldspathic graywacke with rare mafic volcanic rocks.

Throughout the field area the upper contact is with the North East Carry Formation. Although not exposed, this contact is most closely constrained along the northeastern and southeastern shores of Seboomook Lake. Here and elsewhere along the contact, beds on both sides face southeastward and are overturned and indicate that the Ironbound Mountain, at least in the field area, is stratigraphically beneath the North East Carry Formation. This is a reversal of the bedding and stratigraphic relationships which were mapped by Boucot and Heath (1969) in the same area and based on generally poor-quality road pavements. Bedding orientations do not change across the contact and there are no minor structures suggestive of faulting in its vicinity. For these reasons the contact is considered to be a conformable, probably sharp lithologic boundary. In the Caucomgomoc Lake area Pollock (1985) mapped a facies transition between the two lithologies.

To the northeast the rocks of the Ironbound Mountain Formation proper grade laterally into the Grenier Ponds member. Frequency of graywacke beds in the Ironbound Mountain proper lithologically similar to the Grenier Ponds Member increases toward the northeast.

Because the Ironbound Mountain Formation conformably underlies the Gedinian North East Carry Formation it may span the Silurian - Devonian boundary and range from possibly latest Pridolian to earliest Gedinian in age.

### **Grenier Ponds Member**

The Grenier Ponds Member exposed in the extreme northwestern corner of the Penobscot Lake and the adjacent part of the Sandy Bay quadrangles. Additionally, rocks in the northeastern part of the Seboomook Lake quadrangle are assigned to this member. Thickness of the unit is difficult to assess. Since the top of the unit is unknown in the northwest and poorly constrained in the northeast, only a minimum thickness can be estimated. A best estimate is derived from the belt exposed in a small syncline in the northwestern corner of the Penobscot Lake quadrangle (Fig. 2). Here the unit is about 600 m thick.

Massive, light gray weathering, fine- to medium-grained, quartzose lithic wacke comprises about 50% of the member. Fresh exposures are normally medium gray and may be tinted slightly blue or green. Well cleaved dark gray mudstone slate makes up the other half. Wacke thickness ranges between 20 cm and 2 m with an average of about 1 m. Beds are well sorted and either non-graded or poorly graded over their entire thicknesses.

Bedding is planar and regular throughout individual outcrops but beds cannot be traced from one outcrop to another. Both upper and lower contacts with slate are sharp and planar, except where disrupted by cleavage. No primary sedimentary structures were observed in the unit. Cleavage is fairly well developed in about half of the wacke beds while others display none.

Typically the wacke is composed of about 25% quartz, a few percent feldspar, and between 30 and 35% rock fragments. The remainder is micaceous matrix and accessory minerals. Rock fragments are about 75% light-weathering slate and 25% mafic volcanic rocks and siltstone. By imparting a distinct light-flecked appearance to the rock, slate fragments become an important distinguishing characteristic of the Grenier Ponds member.

Interbedded slates are very similar to those of the Ironbound Mountain proper. They are generally between 5 and 20 cm thick and either display no grading or are well graded from siltstone to mudstone. Where grading is not apparent bedding is usually marked by thin silt laminae. Bedding is planar and consistent across individual outcrops but cannot be traced laterally for great distances. Some glacially polished slate outcrops show convolute siltstone laminae. Locally the unit contains thin, matrix-supported conglomerates with clasts of light gray siltstone and slate.

The upper contact of the member was not mapped in the northwestern exposure. In the northeastern exposure the upper contact is with rocks of the Ironbound Mountain Formation proper and is inferred to be a conformable facies transition. Laterally southwestward the graywackes of the Grenier Ponds Member become less prevalent and slate more dominant.

### **Frontenac Formation**

**Background.** Early work on the Frontenac Formation was conducted by McGerrigle (1934) in Quebec Province. He first defined the formation from exposures of sandstone and mafic volcanic rocks southwest of Lac Megantic. On the basis of repetition of volcanic rocks he interpreted the major structure of the region to be anticlinal. McGerrigle found no fossils in the formation and was forced to base his correlation with the Ordovician Ammonoosuc Volcanics of New Hampshire on comparative lithology.

Marleau (1968) worked on the problems of the Frontenac Formation in the Lac Megantic area immediately west of the Penobscot Lake region. His description of the formation is more comprehensive in that he differentiated a number of sedimentary rock types in the formation, including impure quartzites with minor amounts of interbedded slate, fine-grained sandstone, and limestone. Marleau suggested a Devonian age for the Frontenac based on his inference of a regional synclinal structure and lithologic similarity with the Devonian Tarratine Formation then being mapped by Boucot (1961) and Boucot and Heath (1969) in Maine.

Albee and Boudette (1972) conducted a detailed investigation of the bedrock geology of the Attean quadrangle which included work on the pre-Silurian rocks of the BMA and their relationship with overlying Silurian and Devonian units. In corroboration of Marleau's work, they also assigned the Frontenac Formation to the Devonian and suggested that it was stratigraphically above the Lower Devonian North East Carry (Seboomook) Formation. They accepted the synclinal interpretation of Marleau.

The work of Boucot (1961) and Boucot and Heath (1969) defined the limits of the MRS and defined or redefined every formation within it. In considering the Frontenac Formation, Boucot (1961) first indicated that the rocks northwest of the MRS in the Penobscot Lake region were of Cambrian or Ordovician



age. In later reconsideration, Boucot and Heath (1969) designated the rocks to the northwest of the synclinorium as Devonian and loosely correlated the Frontenac Formation with the Tarratine Formation of the synclinorium. However, they indicated that the field evidence was contradictory in nature and that the Frontenac Formation might, at least in part, be older than the North East Carry (Seboomook) Formation.

Chev  (1975, 1983 unpub. mapping) studied the Frontenac Formation and older rocks in the Lac Megantic area of Quebec (Fig. 1). Pollock (1985) studied the Frontenac and North East Carry Formations as well as the pre-Silurian rocks exposed at Caucomgomoc Lake to the northeast of the field area. The consensus of these workers, with which I concur, is that the Frontenac Formation is exposed in an anticlinal structure and that the unit is older than the Seboomook Formation. However, Chev  (pers. commun., 1983) and Bothner and Jahrling (1984) consider the Frontenac to be Ordovician, whereas I consider it to be no older than Silurian or latest Ordovician.

**Description.** The Frontenac Formation is exposed in a structurally complex belt which extends from the vicinity of Allagash Lake southwestward into Quebec Province and northern New Hampshire and northeastern Vermont where it has been mapped in part as the Gile Mountain Formation. Occupying the central portion of the Penobscot Lake region, the belt of Frontenac rocks is about 23 km wide. This width is in part the product of repetition by a series of northwest-verging thrust faults. The belt narrows to about 5 km wide in northern New Hampshire. Because of structural repetition by faulting and folding it is estimated that the outcrop belt represents a six- to eight- fold increase in unit thickness. The unit is probably 3000-4000 m thick.

Most exposures consist of thickly bedded, light bluish-gray to greenish-gray, fine- to medium-grained, graywacke. Graywackes are typically interbedded with thinly bedded dark mudstone slate and are similar in appearance to those of the overlying Ironbound Mountain Formation.

Bedding thickness is an important distinguishing characteristic of the Frontenac Formation. Normally, graywacke beds are between 20 centimeters and 2 m thick. Locally they may reach 3 m or more in thickness, but in some areas may be as little as 10 centimeters thick. Individual beds cannot be traced laterally with confidence from outcrop to outcrop. In outcrops where multiple beds are exposed the bedding thickness may vary through its entire range. In a typical sequence a graywacke bed of a meter or more in thickness is overlain by a few meters of more thinly bedded graywacke topped by thin slate. This may be overlain by a number of thick graywackes where bedding is marked by very thin dark slate layers. Such variation in bedding thickness is not seen in any other unit in the field area.

Few primary sedimentary features exist in the graywackes of the Frontenac Formation. Thick beds most often are massive, well sorted, and either are non-graded or crudely graded over their entire thickness. Upper contacts of graywacke beds are abrupt and usually planar. Locally bed tops may be truncated by broad scour troughs formed during the deposition of the succeeding bed. Cross-bedding, observed in only two outcrops, was not well enough exposed to estimate transport direction. About 30% of the thick beds have poorly defined bedding-parallel laminations. These are usually 1 to 2 millimeters thick, can be traced over outcrop width, and represent a slight decrease in the percentage of matrix material compared with the bulk of the rocks.



In rare occurrences laminations may be arranged in a series of concentric surfaces which are concave-up. Average width of such structures is approximately 10 cm. Convolutions such as these usually are associated with slumping or dewatering events.

Some graywacke beds contain dark gray slate chips 2 to 5 cm in length which are supported by surrounding fine-grained sediment. They are usually found near the base of a bed and gentle warping of some suggests that they were not well indurated when incorporated in the graywacke. Additionally, where cleavage is evident in the bed it cross-cuts the clasts which lack other foliations. This indicates that the slate chips are not from a previously deformed unit and are, therefore, probably intraformational rip-up clasts and disrupted beds.

Most thick graywacke beds of the Frontenac Formation have sharp, planar lower contacts. Basal paleocurrent indicators such as sole markings, flutes, or grooves were not observed in the rocks of the unit. Lack of recognition of these features may be partly due to tectonic overprinting rather than their absence at time of deposition. Throughout most of the unit, widely spaced planar cleavage surfaces are developed. In the graywackes the planes are 5 to 10 millimeters apart and are marked by accumulations of cleavage-parallel platy minerals. Where closely spaced in the wackes the cleavage traces anastomose. At the bases of many thick graywackes, lithologic contrast with underlying slate is great. In these situations, where dip-slip movement on the cleavage surfaces is common, basal structures are obscured by strong lineations resulting from cleavage-bedding intersections. Figure 3 is an example of this type of relationship.

Grain size of the framework constituents of the graywackes varies between 0.10 mm and 0.05 mm, very fine to medium sand size, for most of the formation. At the bases of some of the thicker beds grain size may locally reach 1.0 mm, but few such occurrences were noted in the field. No systematic spatial variation in grain size is evident. Grains are generally subangular in shape although the range, from angular to subrounded, is considerable. Pressure solution is an important grain-shape modifier in a number of samples that show a mosaic texture. Most samples, however, have granular interstitial texture where framework grains are surrounded by finer matrix of platy and very finely granular material.

Framework constituents are quartz (45-55%), plagioclase (0.4-6%), K-feldspar (0.4-15%), and some calcite (0-50%). Rock fragments are found rarely and only near the top of the formation. Matrix is composed of white mica and indistinguishable dark material (5-56%), chlorite (1-15%) and iron-titanium oxides (0-8%). Traces of biotite, clinozoisite, tourmaline, zircon, pyrite, ankerite, actinolite, and detrital mica are found in many samples. The percentage of unidentifiable matrix material generally decreases while the amount of chlorite and white mica increases in zones of contact metamorphism. Calcite quantities vary highly throughout the formation and with little consistency in distribution. Pyrite often imparts a rusty or rust-spotted appearance to the unit where it is prevalent.

Mudstone slate beds in the Frontenac Formation may be easily mistaken for those of the Ironbound Mountain especially where slate is the only lithology exposed in an outcrop. This is a particular problem near the Frontenac -





Fig. 3. Sketch of cleavage-disrupted bedding in the Frontenac Formation. Slate bed is stippled and graywacke unpatterned except for cleavage traces. Arrow indicates topping in wacke. Lens cap for scale.

Grenier Ponds contact in the northwest. There slate beds are more frequent in the Frontenac and are often poorly exposed in logging roads. Throughout most of the Frontenac, slate is only a few centimeters thick where it is interbedded with thick sandstone. Color of the slate varies between dark gray to olive-green gray. The greener variety results from increased chlorite content in the vicinity of prevalent mafic intrusions, which increase in frequency southwestward. Toward the top of the formation the interbedded slates generally become thicker and some beds may be silty. All slates are well cleaved and the siltier beds are well graded.

#### Canada Falls Member of the Frontenac Formation

Volcanic rocks of this member crop out in one major northeast-trending, lens-shaped body and two minor lens-shaped bodies (Fig. 2). The smaller bodies are partially fault-bounded. Structural repetition by faulting and folding increases the width of the principal belt 3 to 4 times over the estimated stratigraphic thickness of 1000 to 2000 meters.

Rocks of this member are generally green basalt which weathers light green to a depth of a centimeter. Exposures can be categorized as well-pillowed and poorly cleaved, or poorly pillowed and well-cleaved. Pillowed sections are more prevalent toward the top of the section and represent about 40% of the outcrops examined.

Where well developed, pillows range in diameter from 10 centimeters to 2 meters. "Bread-loaf" tops of many provide excellent topping evidence. Propylitized rims and interstitial material, which weather yellowish-green and locally maroon, give way to medium-grained texture at the cores of many pillows. The core texture is similar to, and easily confused with, the texture of hypabyssal mafic rocks found in the area. Small knots (< 10 cm diameter) of milky white quartz and pistachio-green epidote are common in interstitial areas but rarely found inside pillows. Centers of large pillows are jointed in a crudely developed radial pattern.

Many pillows contain vesicles and calcite spherulites which attain a diameter of 1 cm at their largest. Spherulites and vesicles rarely exceed 10% of the surface area of the rocks, while the maximum is as much as 50% of the rock. Approximately 20% of the pillows have these features.

In outcrops which are not pillowed, cleavage is the most obvious structure. It is generally penetrative, spaced only a few millimeters apart, and planar. Outcrops of this type are often monotonously uniform both in medium grain size and texture over the entire exposure except for meter-sized lenses of uncleaved, poorly jointed rock. These lenses are similar in appearance to the jointed cores of some pillows and may comprise 10% of the outcrop.

The mineralogy of the basalt is the product of significant alteration of the primary minerals. Grain size is generally too fine to permit identification of minerals in hand specimen, except 20% of the rocks which contain white, euhedral phenocrysts of plagioclase. Phenocrysts are up to 2 millimeters in length. Medium-grained samples exhibit a felted groundmass of lath-like feldspar and clustered chlorite blades. In thin section the results of alteration are apparent in the replacement of primary mafic constituents by chlorite and epidote. Plagioclase has been completely albitized and contains inclusions of calcite and epidote. Magnetite is a common opaque mineral.

Other lithologies constitute a minor percentage of the Canada Falls Member. Most notable is a thin pyroclastic layer of volcanic bombs and tuff located near the base of the unit. The layer is only exposed in one locality and consists of light brown, elongate, highly vesicular clasts imbedded in dark gray, foliated tuffaceous material. They are completely supported in the very fine-grained uniform tuffaceous matrix of aquagene(?) origin.

Additionally, the member contains minor beds of intercalated siltstone and mudstone slate. These generally drape over the irregular upper surface of the pillow basalt and are only a few meters thick.

### **Contacts of the Frontenac**

Contact relationships of the Frontenac Formation with the overlying Iron-bound Mountain Formation are central to establishing its stratigraphic position and age. Of importance in determining the nature of the contact are not only lithologic and sedimentary factors but structural ones as well. Lithologies of both units in the vicinity of the contact are similar. Most Frontenac beds in the northwestern contact zone are thick graywacke. Across the area between the Marie Petuche Fault (Fig. 2) and the contact zone, progressively thinner graywacke beds and thicker slate sections are interbedded. Slate is particularly prevalent in road exposures and between hummocks under-



lain by graywacke. Across the contact in the Grenier Ponds Member, the proportion of wacke decreases to about 50% and individual beds are thinner. Slate becomes proportionally more important.

## STRUCTURAL GEOLOGY

The dominant structure of the Penobscot Lake region is interpreted to be a broad, northeast-trending anticline situated near the southeastern margin of the CVGS. This anticline has been sliced by imbricate, northwest-directed thrust faults (Fig. 2), the most continuous of which, the Marie Petuche fault, is nearly coincident with the anticlinal axis. Southeast of this fault most of the strata face southeastward. Northwest of the fault, beds are mostly sub-horizontal or northwest-facing. This suite of thrust faults is interpreted as a positive flower structure related to right-lateral movement on the Thrasher Peaks fault, which traverses the area just south of the Penobscot Lake region (Fig. 1) (see Marvinney (in press) for detailed discussion of strike-slip faulting in this region).

### Cleavage

A number of cleavages are recognized in the field area, the most spatially widespread of which is a northeast-trending, steeply dipping, penetrative cleavage, here termed regional cleavage ( $S_1$ ). This is best developed in the fine-grained rocks of the Seboomook Group, especially the Ironbound Mountain Formation. In lithologically homogeneous outcrops of this formation, the cleavage is characterized by remarkably parallel, planar seams or domains which impart a fissile character to the rock. Thin-sectioning reveals the pervasive and penetrative cleavage to be defined by preferred orientation of phyllosilicate and elongated grains. Cleavage intersects bedding at an oblique angle and the spacing between seams in outcrops appears to be less than one millimeter. This cleavage would typically be termed slaty in fine-grained rocks.

When the outcrops consist of interbedded graywacke and siltstone or mudstone, as they commonly do in the Frontenac Formation, the cleavage takes on a different appearance. Cleavage seams refract sharply as they cross sharp contacts between the bases of graywacke beds and tops of finer-grained beds. Upward through graded beds they refract more gradually. Concomitant with refraction is a decrease in seam spacing. In outcrops of coarser grained rocks the cleavage appears fracture-like and is spaced by as much as one centimeter. Cleavage seams anastomose and, in spite of their fractured appearance, most often are defined by concentrated opaque materials, oriented phyllosilicates, and partially by dimensional preferred orientation of elongate grains.

The regional cleavage ( $S_1$ ) is the earliest generation of cleavage in the area and is roughly axial planar to small folds. Its style and orientation is similar to cleavage attributed to the Acadian orogeny in other areas of the northern Appalachians (Osberg, 1978).

**Orientation of  $S_1$ .** The field area was divided into eight structural domains for the purpose of comparing cleavage orientations. Domains were selected to elucidate changes in structural grain inherent to (1) location, and (2) units of differing age. Stereoplots of poles to  $S_1$  for each domain are shown in Figure 4.

Cleavage orientations are noticeably different in the southwest and northeast parts of the region, respectively. In domain I, underlain mostly by the Frontenac Formation, cleavage has a strong NE trend and is nearly vertical, on average, judging from the symmetry of the plot. In structural domains II and III the symmetry is not as well developed. More cleavage planes dip to the southeast and they show more variation in strike azimuth. The strong east-west component in the pole diagram indicates that a significant number of planes strike north-south. Similar trends are seen in domain IV but they reach their maximum expression in domain VI, the most northeastern. The stereoplot here is bimodal, showing that almost as many planes are oriented north-south as northeast-southwest. A similar, although subdued progression is seen through domains V, VII, and VIII. However, many more cleavage planes in these southern domains dip to the northwest. In general, cleavage strikes more toward the north in the northeast and dips northwestward in the south.

**Fault-related cleavage.** A later cleavage ( $S_2$ ) is related to thrust faults. These are distinguished from  $S_1$  first by their appearance, often brittle, and second, by their cross-cutting relationships and local nature. In appearance  $S_2$  contrasts strongly with  $S_1$ , at most localities being a true fracture cleavage, following the definition of Hobbs et al. (1976). Fractures are a

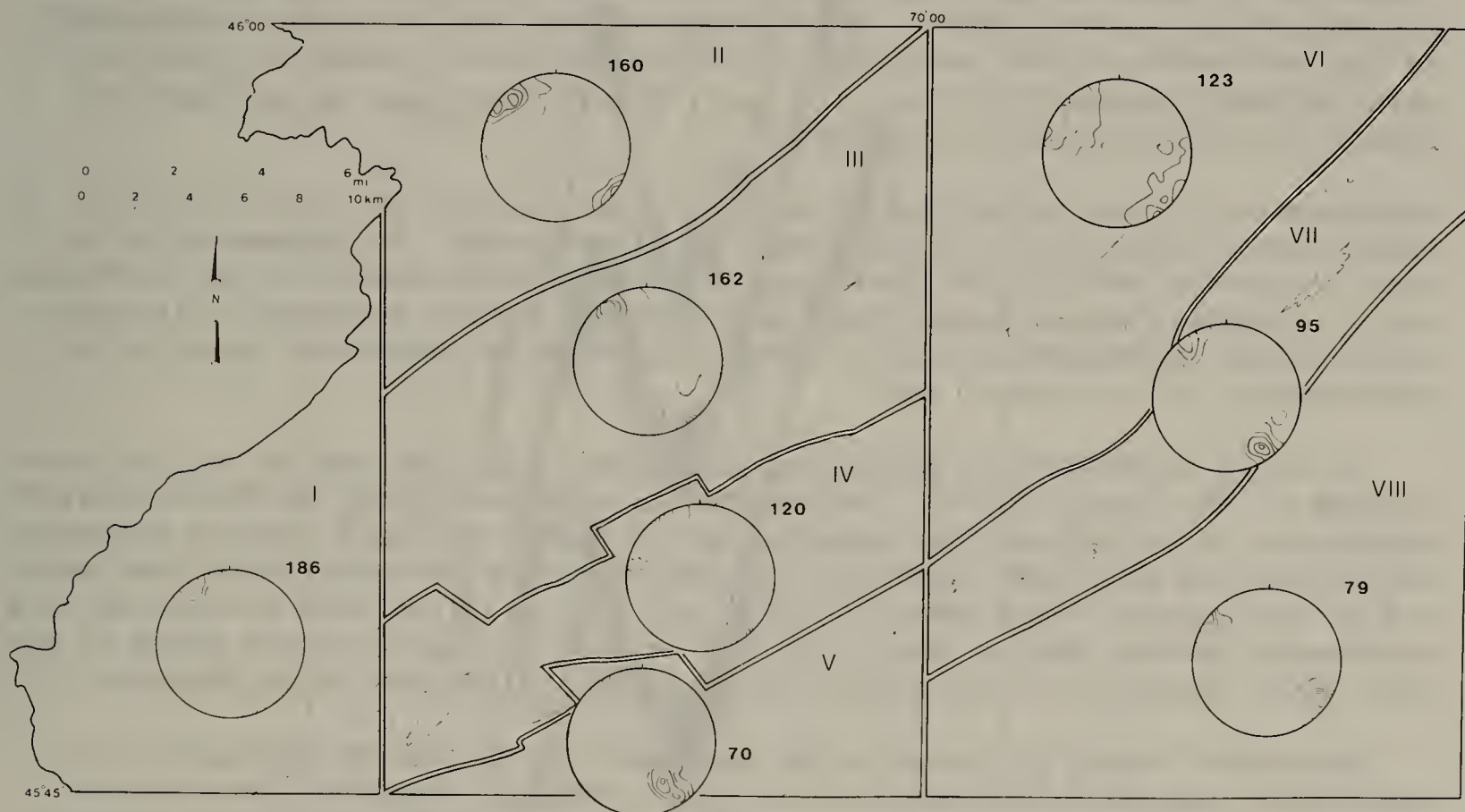


Fig. 4. Lower hemisphere, contoured stereoplots of poles to  $S_1$  in 8 domains within the field area. Number of poles given for each plot.



few centimeters apart and nearly vertical. The intersections of  $S_1$  and  $S_2$  form microlithons which break away to leave very craggy outcrops. Similar structures in the Helvetic Alps have been termed pencil cleavages by Ramsay (1981), where they are unambiguously related to thrust and nappe development.

Other examples of fault-related cleavage clearly show that it cross-cuts both bedding and  $S_1$ . Figure 5 shows the base of a near-vertical graywacke bed of the Frontenac Formation, on the left, and part of an underlying siltstone slate bed on the right. The bed base shows broadly rounded features which are interpreted as major load structures that have been accentuated by cleavage-parallel slip. Regional cleavage is well-developed in both lithologies and is sub-horizontal. This attitude is unique and indicates that post-cleavage rotation toward the northwest has occurred. Close examination of regional cleavage traces in the area where slate is pinched between sand lobes reveals a sub-vertical crenulation cleavage. The orientation and associations are believed to be related to thrust rotation of regional cleavage toward the northwest.

## Folds

**Small-scale.** Outcrop-scale folds are often encountered in the field area. The majority are tight to isoclinal, show some minor thickening in the hinge zones, and have amplitudes ranging from a few centimeters to several meters (Fig. 6). Most folds verge to the northwest, but some in the southern part of the area are southeast-verging. These are asymmetric folds in which long limbs dip at moderate to steep angles toward the southeast, and short limbs dip moderately northwestward. Axes plunge at shallow to moderate angles both to the northeast and the southwest. Most of the axial planes are inclined with regional cleavage ( $S_1$ ) roughly axial planar, but some in proximity to thrust faults are close to recumbent.

**Large-scale.** The orientations of bedding planes across the area mimics, in a broad sense, the style of folding seen in minor folds. The asymmetry of folding is clearly seen in the dominance of beds dipping steeply to the southeast and a relatively minor number dipping moderately to the northwest. In particular in the southeastern belt of Ironbound Mountain Formation, there is a predominance of overturned beds.

In areas of relatively continuous exposure it is possible to map the axes of some of the larger folds. The northwestern extremity of the field area, in particular, provides such an opportunity. A number of small, doubly plunging anticlines and synclines which are overturned to the northwest have been mapped in the Grenier Ponds Member. These may have resulted from overriding of a northwest-directed thrust sheet. In the central and southeastern parts of the area where thrusting is important, folds of this size tend to be obscured.

Northwest facing of units in the northwest and southeast facing in the southeast, together with repetition of the Ironbound Mountain Formation on both sides of the Frontenac Formation (Fig. 2) suggest that the main structure in the center of the field area is a broad anticline. This structure is complicated by thrust faulting.



Fig. 5. Sketch of cleavage relationships in thrust area. Sub-horizontal  $S_1$  in sandstone bed topping left (NW) is crenulated by near-vertical  $S_2$ .

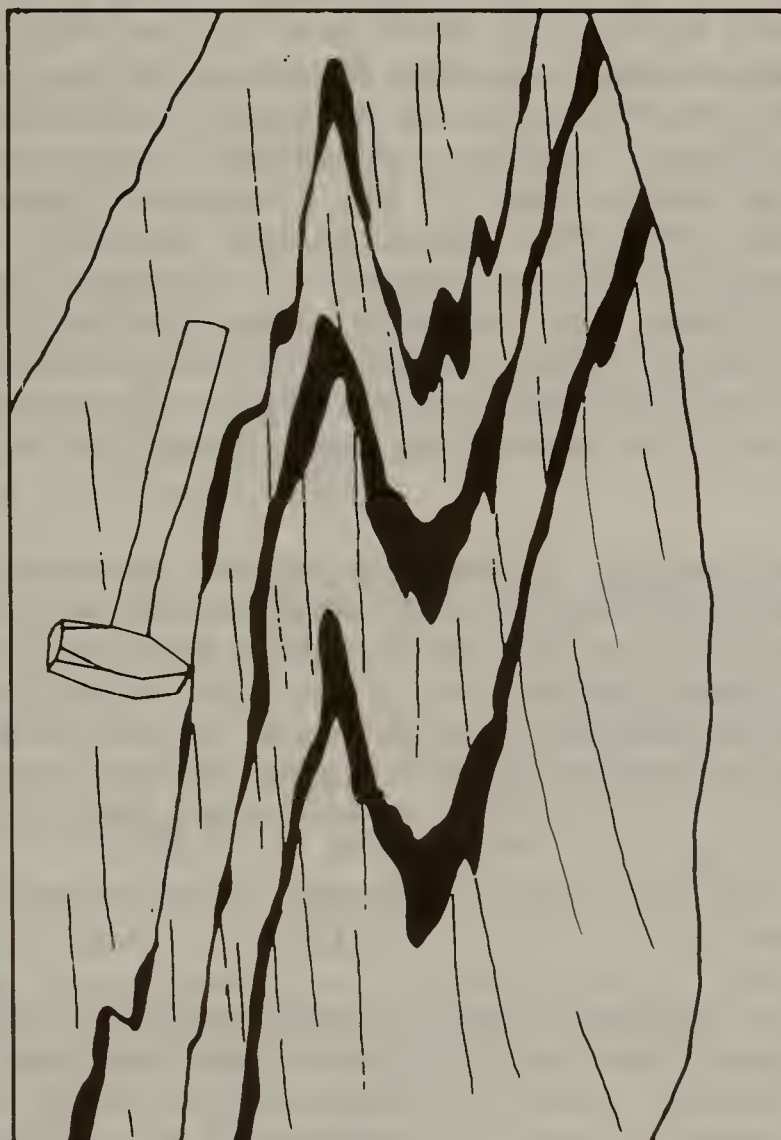


Fig. 6. Sketch of typical NW-verging fold in Frontenac.



## Thrust Faults

Within the main part of the Penobscot Lake region the most important mode of faulting is northwest-directed, low- to moderate-angle thrusting. Although three separate faults were mapped, two of these may be imbrications of the first and longest, the Marie Petuche fault, named for a pond in the Penobscot Lake quadrangle which it traverses. Occurring entirely within the Frontenac Formation, this fault extends for more than 25 km from the northeast corner of the Penobscot Lake quadrangle through the international border station on U.S. Route 201 in the Sandy Bay quadrangle, where it is superbly exposed in artificial road cuts. Its extent beyond these areas is unknown.

At the international border station on U.S. Route 201, the Marie Petuche fault cuts thick graywacke and interbedded slate of the Frontenac Formation. Through much of the outcrop the dominant structures are bedding plane partings or separations, and other fractures developed at a low angle to bedding. In combination with a number of cleavages, the partings and fractures impart an extremely brittle appearance to the outcrop. Also, in parts of the outcrop individual beds are segmented by southeast-dipping shears, which makes bedding extremely difficult to trace.

The Marie Petuche fault is mapped northeastward along the international border and the southeast side of Green Mountain for 3 reasons. This is coincident with a narrow band of kinking which may be fault-related as it is a narrow, linear zone involving localized strong, nearly horizontal shortening. Other fracture cleavages similar to those seen at the border station exist along this trend. Most importantly, the structural style and stratigraphy abruptly change across the zone. On the southeast minor folds verge toward the northwest and are tight. Beds face southeast; lower stratigraphic levels are exposed towards the fault. Beds in the Frontenac Formation are thin to medium in thickness and interbedded slate is less prevalent. Northwest of the fault, folds are generally broad and upright or southeast verging. Beds face northwestward; the Frontenac - Ironbound Mountain contact is approached in a short stratigraphic distance. Beds in the Frontenac Formation are thick, and interbedded slate is more frequently encountered. All observations suggest an important discontinuity at the zone with some repetition of section in the Frontenac.

In two other areas faulting of the same nature has resulted in stratigraphic repetition of the section. One occurrence is in the central part of the Penobscot Lake quadrangle (Fig. 2), where a thin wedge of Ironbound Mountain Formation is in depositional contact with the underlying Frontenac Formation on the northwest and structurally beneath it on the southeast. All units dip and face southeastward indicating an older-over-younger relationship. Extensive kinking and fracture cleavage are developed in the Ironbound Mountain Formation. Because the style of deformation is similar to that seen along the Marie Petuche fault, this fault is interpreted to be an imbrication of northwest-directed thrusting.

The second fault is inferred along the northwestern limit of the Canada Falls Member where it is in contact with excellent mudstone slate of the Ironbound Mountain Formation. Bedding in the latter faces southeastward into the volcanic rocks in which it was not possible to determine bedding attitude near the contact. The fault trace is represented by a narrow topographic low in which there is no outcrop.

## High-angle faults

A number of high-angle fault sets offset some of the stratigraphy and earlier structures in the region. One or possibly two sets of northeast-trending faults of near-vertical dip and one set of northwest-trending "cross" faults have been delimited.

A high density of northeast-trending faults exist in the southwest part of the field area, although they are not limited to this part of the area. They are known from repetitive juxtapositions of stratigraphic units on the southeast flank of the mountain and from road cuts along U.S. Route 201 described by Albee and Boudette (1972). Faults there are mapped as having normal dip-slip separation because the younger Ironbound Mountain Formation is juxtaposed against the Frontenac in an area largely underlain by the latter.

Three northeast-trending "cross" faults offset the strike belt of the Ironbound Mountain Formation in the southwestern part of the area (Fig. 2). Their extent beyond this belt is not well defined and is based on prominent topographic lineaments and possible offsets in stratigraphic units. Although the map patterns at first inspection suggest strike-slip displacements along these faults, primarily, they are more likely related to other late, northwest-trending faults throughout northern Maine (Osberg et al., 1985) which are primarily dip-slip.

## SUMMARY

The stratigraphic units of the Penobscot Lake region are interpreted as having been deposited in a marginal basin which may have been formed, at least in part, by transtensional processes. Figure 7 reviews the geologic settings of deposition of these units. Rocks of the Frontenac Formation are interpreted as proximal turbidites representative of fairly steep basin slopes or proximity to source areas. The general lack of volcanic rock fragments in Frontenac rocks argues against a pure back-arc setting, at least in terms of the classification scheme of Dickinson et al. (1983). The mafic volcanic rocks of the Canada Falls Member are suggestive of an extensional setting although chemical analyses are ambiguous as to continental or oceanic affinity for these basalts (Marvinney, 1986).

The occurrence of thick bedded, proximal turbidites and frequent mafic sills, as seen in the southwestern part of the area, constitutes an example of a sediment-sill complex as described by Einsele (1985) in the central Gulf of California. The high flux of sediment into this narrow spreading basin hampers the upward movement of magma at the spreading center resulting in the intrusion of magma as sills rather than extrusion as pillow basalts and flows seen in typical spreading centers.

Although also turbiditic, the more distal nature of the deposits of the Ironbound Mountain Formation indicate either an increase in water depth, reduction in source area relief, or a combination of the two. At this time the basin in which the Frontenac was deposited may have amalgamated with other basins in the region such that the units of Early Devonian age are fairly similar in lithology throughout the area. The coarser grained sedimentary rocks of the North East Carry Formation are indicative of the development of mountainous areas toward the southeast as the Acadian orogeny was initiated.



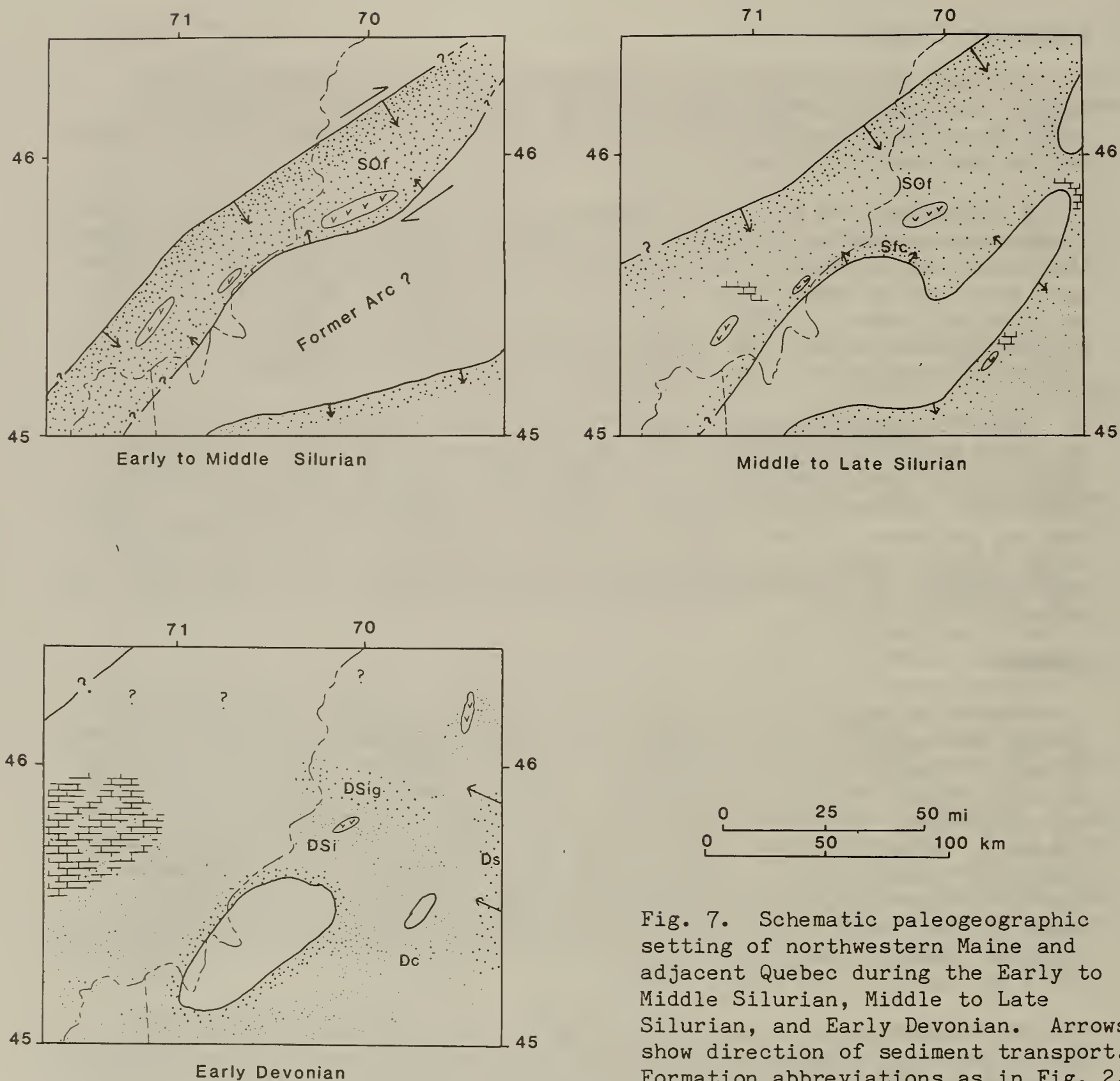


Fig. 7. Schematic paleogeographic setting of northwestern Maine and adjacent Quebec during the Early to Middle Silurian, Middle to Late Silurian, and Early Devonian. Arrows show direction of sediment transport. Formation abbreviations as in Fig. 2.

The structures of the region are Acadian in origin. An interpretive cross section is shown in Figure 8. This shows that the rock units generally young southeastward and that they are cut by a series of imbricate, northwest directed thrust faults. These are interpreted as forming part of a "positive" flower structure about the right-lateral Thrasher Peaks fault, similar to occurrences noted elsewhere by Harding (1974) and Harding and Lowell (1979).

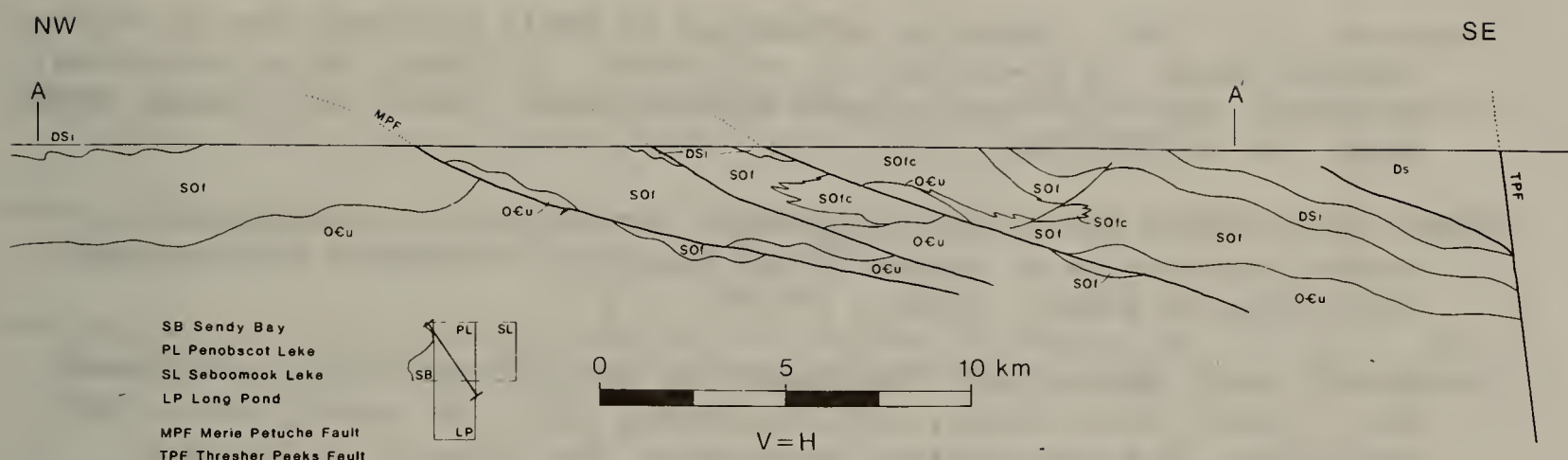


Fig. 8. Interpretive cross section from NW to SE across the Penobscot Lake quadrangle.

Back-thrusting may be responsible for juxtaposing Frontenac and Ironbound Mountain rocks in the southeast and for overturning part of the section there.

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#### ITINERARY

The trip will begin at the parking lot of St. Joseph's Catholic Church on Rte. 15 in Rockwood. We will be on dirt roads all day and there will be no opportunity to purchase provisions once under way. Please note that these are private paper company roads and that right of way should be yielded to all logging trucks. At each stop be sure to park vehicles as far off the roadway as possible. Unless there is alot of rain, the roads will be dusty.

#### Mileage

#### Total Int.

0.0		St. Joseph's Catholic Church. Turn left (west) on Route 15.
0.1	0.1	Turn right onto bridge crossing Moose River. Bear right at north end of bridge.
0.6	0.5	Enter Great Northern Paper Company land.
8.8	8.2	Intersection. Proceed straight ahead.
11.6	2.8	Road intersecting from right. Keep left.
12.7	1.1	<b>STOP 1.</b> Park on the left side of the road on broad shoulder before the road curves to the right. Typical blue-gray graywackes of the North East Carry Formation are best exposed in the road cut on the east side of the road in the curve. Bedding in the graywacke here ranges from 2 - 40 cm in thickness, with an average around 10 cm. Although there is variability



in bed thickness, there are relatively few thick beds and few thin beds. Each bed is well graded from fine grain size at a very sharp base, smoothly upward through mudstone at the top. The average percentage of sandstone vs. mudstone across the outcrop is about 50-50.

Bedding is oriented N59E 72NW and is overturned.  $S_1$  is a fracture cleavage in sandier sections which grades into slaty cleavage in the mudstone sections. Typical refraction of cleavage is well demonstrated here.  $S_1$  is oriented N57E 56NW in slate sections.

Proceed north on dirt road.

- 15.3      2.6      **STOP 2.** Park on right at Mile Marker 87. Walk east about 200 m through slash to the south end of the small pond that can be seen from the road. This massive, light gray-weathering, fine- to medium-grained graywacke is typical of the amalgamated graywacke beds found locally in the main body of the Ironbound Mountain Formation. This lithology is more typical of the Grenier Ponds Member into which the main body of the formation grades to the northeast.

No clear evidence of bedding can be seen in this 5 m thick outcrop. Some layers containing large (5-10 cm) dark gray slate chips are evident, especially in the outlet stream from the pond, but their orientation is difficult to establish. Cleavage runs from the graywacke through the slate chips suggesting that these were undeformed rip-up clasts when the graywacke was deposited.

Cleavage is much more pervasively developed in these graywackes than in the fine-grained sandstones of the North East Carry Formation seen in the previous outcrop. This is due to a greater percentage of matrix micas in these graywackes and is typical of the appearance of  $S_1$  in the coarser grained rocks of the Ironbound Mountain Formation and the Frontenac Formation.  $S_1$  is oriented approx. N55E 73NW.

Proceed northward on dirt road.

- 15.8      0.5      **STOP 3.** Park as tightly together as possible on the right side of the curve. The outcrop on the left shows atypically poorly bedded siltstone and mudstone slate of the Ironbound Mountain Formation. This is one of the few debris flow or olistostromal sections in the formation. Small (2-3 cm) clasts of graywacke and slate can be found in the massive section at the south end of the outcrop. These are most easily seen on the naturally weathered upper surface of the outcrop. On the north side of the outcrop, a thick, more-or-less in place, poorly cleaved graywacke bed consisting of a coarser grained, less micaceous rock than seen in Stop 2, is oriented N59E 60SE.

Cleavage is more poorly developed here than in most parts of the Ironbound Mountain Formation owing to a greater silt content here.  $S_1$  is oriented N59E 80NW.

Proceed north on dirt road.

- 16.2      0.4      **STOP 4.** Park in pull-out on left side of road. Outcrop is on right side of road. This is the typical slate of the Ironbound Mountain Formation and constitutes 95% of the main body of the formation. The dark gray siltstone and mudstone slate has very regular bedding and well developed cleavage. Beds are generally 2-15 cm thick, have sharp bases, are well graded from siltstone to mudstone in the thicker beds, and are upright, topping to the southeast. Some fine laminae also mark bedding which is oriented is N63E 48SE.

Cleavage is penetrative and very planar here as it is throughout the slate of the Ironbound Mountain. It is oriented N58E 79SE.

The contrast between these rocks and those of the North East Carry Formation are the regularity of the slate and the predominance of siltstone and mudstone over any coarser grained beds.

Continue north on dirt road. The contact with the Frontenac Formation is crossed within 0.2 miles.

- 16.6      0.4      Great Northern Paper Co. gate.
- 16.8      0.2      Intersection. Continue straight. Road to right leads to Seboomook Dam.
- 16.9      0.1      **STOP 5.** Park as close together as possible on the right side of the road. Walk east (right) 25 m down the slope toward the small stream to where the slope steepens markedly. Walk approximately south along the slope break to large outcrops of massive graywacke typical of the Frontenac Formation. The graywackes here are several meters thick, poorly graded, mildly calcareous, light gray-weathering, fine-grained rocks. The graywacke is very micaceous and, as a result, well cleaved. Bedding is elusive. Only where interbedded dark gray slate 10-20 cm in thickness is found can bedding orientation be established. It is oriented N48E 71SE and is upright.  $S_1$  is oriented N58E 82SE.

Continue north on dirt road.

- 17.7      0.8      Intersection. Keep right toward Windy Pitch.
- 19.0      1.3      Intersection. Keep left toward Pittston Farm.
- 19.2      0.2      **STOP 6.** Park at the south end of causeway crossing the western end of Seboomook Lake and walk 100 m northeast across



open area to large promontory. These are pillow basalts of the Canada Falls Member of the Frontenac Formation. At the south end of the outcrop, thinly bedded (2-5 cm), brown siltstone overlies green basalt and in-fills the irregular surface of the basalt. This siltstone lithology is uncharacteristic of the Frontenac or Ironbound Mountain and may represent a local depositional environment.

Pillows in the basalt are well developed and variable in size (a few cm to 1 m long). They are very elongate here although elsewhere they can be quite rounded. In spite of the elongation, topping can be determined from the rounding of tops and necking of bases of pillows. The beds are nearly vertical and top to the SE.

On the northern side of the outcrop is a sequence of thin, buff-weathering, very fine grained arenites. These are composed primarily of quartz and feldspars with very little matrix. The former matrix material has been metamorphosed by the overriding basalt to form green biotite and some actinolite.  $^{40}\text{Ar}/^{39}\text{Ar}$  dating of the biotite is inconclusive.

$S_1$  is very poorly developed in the main part of the basalt. Toward the north side the basalt becomes more massive and better cleaved. In many pillows, the coarser grained cores are jointed into a very blocky appearance. In the non-wave worn parts of the outcrop, this feature is almost the only indication of pillowing.

Continue across causeway. Outcrops between here and the next stop will be basalt.

- |      |     |   |
|------|-----|---|
| 19.7 | 0.5 | Intersection. Continue straight.  |
| 20.0 | 0.3 | Bridge over the South Branch Penobscot River. Pittston Farm Ranger Station on left.   |
| 20.4 | 0.4 | Intersection. Turn left toward Canada Falls Dam. The road winds its way toward the dam, affording several excellent views of the river. The rocks are all volcanics of the Canada Falls Member.   |
| 23.1 | 2.7 | <b>STOP 7.</b> Canada Falls Dam. Park in the open area near the top of the dam. The small island directly across the log boom is underlain with basalt. The island farther southwest is underlain by slate of the Ironbound Mountain. Toward the northwest, basalt is encountered within 100 m. The slate is probably exposed in a fault sliver. Evidence for faulting is apparent in the rocks exposed along the base of the dam toward the river. The slate is silty Ironbound Mountain. These are contorted and shot through with quartz veins including one massive meter-thick vein at water's edge. These are thought to be the consequence of recrystallization of quartz in a shear zone. |

The outcrop which juts out downstream is well bedded and well cleaved slate of the Ironbound Mountain. Bedding is oriented N47E 88NW, and upright.  $S_1$  is oriented N59E 80SE. A late fracture cleavage related to faulting is oriented N27E 75NW.

Drive back down the Canada Falls Road to the main road.

- 25.8      2.7      Turn left onto main road.
- 26.6      0.8      We have crossed one thrust fault and the rocks about 150 m through the woods to the left at the slope break are graywackes of the Frontenac Formation which face to the southeast. Do not stop.
- 27.4      0.8      Intersection. Keep right.
- 28.0      0.6      **STOP 8.** Park on the right just before the road steepens sharply. Walk down the slope to the right about 50 m to the North Branch Penobscot River. At water's edge is one outcrop of rocks thought to be transitional between the Ironbound Mountain and the Frontenac. The well bedded, well graded light blue-gray graywacke and dark slate are overturned to the northwest (N63E 81NW). Beds are 10-20 cm thick and very disrupted with slate injections along the bases of many of the coarser beds. There is considerable convolute bedding and minor folding. The graywacke is very micaceous and rust-spotted, reminiscent of the Frontenac Formation although the grading and laminations are uncharacteristic of that formation. The overturning is the consequence of southeast-over-northwest thrusting or reverse faulting.

Continue northeast on dirt road.

- 29.0      1.0      Intersection. Bear right.
- 29.4      0.4      **STOP 9.** Bridge over Leadbetter Falls. Cross the bridge and park on the left side of road well away from the bridge. These rocks are transitional in character between Ironbound Mountain and Frontenac lithology. The character of the rocks is sandier than most of the Ironbound, but alot more variable in mica content and brown color than the North East Carry. There is generally more slate than in the Frontenac but frequency of Frontenac-like wacke increases down-section (upstream). The structural setting here is a large, asymmetrical, NW-verging fold with a wavelength of several hundred meters. On the south side of the bridge, bedding is regular and is oriented N5E 10E. Beneath the bridge bedding is sub-horizontal. North of the bridge bedding abruptly dips steeply NW, and considerable slip and shear along cleavage disaggregates bedding. This NW dipping limb of the fold has minor folds which verge to the SE. On the west side of the river, continuing upstream, bedding begins to dip SE again. Minor folds verge NW, are sheared SE-over-NW, and show about 50% shortening. About 200 m upstream from the bridge, bedding once again dips gently to moderately SE.



Continue east on dirt road.

30.3      0.9      Intersection with the Golden Road. Turn right (SE) and watch for trucks!!!

31.9      1.6      **STOP 10.** Pull as far off the road on the right side as possible. This stop is about 200 m south of a dirt road marked High Road. Cross the road to low outcrops on the E side. This shows graywackes of the Frontenac and a good illustration of crenulation cleavage ( $S_2$ ) related to faulting. Bedding is nearly vertical and facing NW. The disrupted base of a fine-grained wacke shows  $S_1$  to be dipping gently to the SE, a marked change from the orientation noted thus far. Where dark slate is injected into the base of the wacke, a near-vertical crenulation cleavage ( $S_2$ ) has developed, possibly in response to northwest-directed thrusting. This outcrop (illustrated in Fig. 5) is thought to be on the lower plate of a thrust.

Continue SE on Golden Road.

32.5      0.6      **STOP 11.** Park on the right. More Frontenac Formation, facing SE in the upper plate of a thrust. Beds are generally 40-50 cm thick and poorly graded. Cleavage is near vertical but strongly refracted as the beds become finer grained. Minor folds verge NW and there is still evidence of brittle deformation.

Turn around and head north on the Golden Road.

34.7      2.2      Intersection with Pittston Farm road. Continue straight. Small outcrops along the road and on the west side of the river show that we are continuing down section.

38.9      4.2      **STOP 12.** Large outcrop of Frontenac Formation on right. This outcrop shows some of the variability of bedding thickness in the Frontenac, ranging from several meters down to a few cm. Except for one massive graywacke, the unit is much siltier and slatier than elsewhere. This is near the top of the formation near the contact with the Ironbound Mountain and the rocks on the north side of the outcrop are most characteristic of this transition.

The asymmetry of folds indicate vergence to the SE in this lower plate of a thrust. There are several near-vertical fractures marked by quartz veining which are related to thrusting. On the north side, bedding dips very gently which is characteristic of the entire northern part of the field area.  $S_1$  is oriented N43E 84SE. Fold axis trend N38E and plunge 9E.

**END**

Turn around and head south on Golden Road. Turn right off Golden Road at mile 77. Cross North Branch and bear left at 3 more intersections. Pass ranger station, cross South Branch, continue straight across causeway on Seboomook Lake. Bear right at next intersection and straight at following ones to Great Northern gate. Follow the road straight (S) through all intersections to Rte 15. Turn left to Greenville.

# DATE DUE

JUN 18 1993

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